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TILL FABRIC

BY CHAUNCEY D. HOLMES

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ABSTRACT

A study of the arrangement of component materials in undisturbed till, the till fabric, shows that at most localities the imbedded stones tend statistically to lie so that their long axes are parallel to the direction of glacier flow at the time of deposition. In a few localities the dominant statistical preference is for alignment at right angles to that direction. Presumably the parallel orientation was normally acquired by sliding, and the transverse orientation by rotation, and permanent deposition commonly occurred without loss of alignment. Fabric analyses indicate that stones of certain forms and degrees of roundness (enumerated in text) have a greater-than-average statistical chance for deposition either parallel to, or transverse to, the direction of transport. Such stones thus serve as guides to the direction of glacier flow and are independent of other evidence. Characteristic depositional attitudes of certain types of till stones permit inferences regarding the probable nature of the transportational environment.

INTRODUCTION

FACT OF TILL ORGANIZATION

Undisturbed till has an inherent organization. At most localities this organization manifests itself in the tendency of imbedded stones to lie so that their longest dimension or axis coincides approximately with the direction of glacier flow at the time of deposition. Such tendency is best revealed by a statistical study of the positions of at least 100 till stones from any one locality. The results can be expressed by diagrams such as the simple "rose" figure or by the contoured diagram commonly used in petrofabric studies. This makes possible the determination of ice-flow direction at places where no striae or other criteria are available. Moreover, the direction of glacier flow is known to have varied from time to time at any one place. Hence, in problems such as locating mineral deposits by tracing drift fragments to their sources (Sauramo, 1924), data from till-stone orientation may be as essential as those from striae on bedrock beneath the till.

HISTORICAL STATEMENT

Miller (1884) published probably the first critical observations on till-stone orientation. In describing "pavement boulders" [1] in the till near Edinburgh, Scotland, he stated (p. 167):

"The longer axis of the stone is often directed in the line of glaciation, and the pointed end is frequently, but not always, toward the ice."

Smaller stones may be oriented in "fluxion structures" around the larger boulders.

"It does not follow, however, that wherever we find an orientation of boulders in the till there was fluidal motion in the layer in which they lie. If the ice had a fluxion structure of its own, such boulders as were incorporated within its mass would arrange their axes conformably; and, when they lagged and came to rest and were imbedded, they might retain in many cases the arrangement that marked them when in motion" (p. 187).

[1] Stones with strongly striated upper surfaces, generally occurring in definite horizons in till and constituting a "pavement" on which the superjacent till rests.

Bell (1888b) carried the investigation further by visiting a number of Swiss glaciers to determine whether such orientation of stones existed in transit. He concluded (p. 341)

> ". . . that the tendency of boulders on all glaciers is to assume a longitudial position, and that this is most observable . . . on large glaciers, where the obstructions are fewer in proportion to the mass, and produce the least disturbing effect."

Unfortunately, both Miller's and Bell's data are largely qualitative and selective.

Upham (1891) described the characteristic position of "oblong" stones in subglacial till as having their long axes parallel to contiguous striae. Flat stones were said to lie parallel with the surface of deposition. However, Upham recorded no systematic investigation.

The results achieved by these early investigators failed to attract the notice they deserved. James Geikie's classic treatise (1895, p. 15, 62) probably reflects the prevailing attitude of that time. Although mentioning that "in certain regions the large and small till stones are oriented in parallel fashion," and citing Miller's work, Geikie described till as a clay containing a "confused and pell-mell mixture of stones." For many years this concept of chaotic agglomeration remained practically unchallenged. Twenhofel (1932, p. 86) summarized more recent opinion by stating that ground moraine consists of "unstratified and unorganized material."

Papers by Richter (1932, 1933, 1936) are among the most important published works of recent date. His discoveries apparently developed as a modification of the method used by the Scottish geologists for determining the direction of glacier flow from striae on the upper surfaces of "pavement boulders" in areas where striated bedrock ledges are absent. In northern Germany, Richter found that the long axes of elongate stones tend to parallel the direction of pavement-boulder striae, even though such stones themselves bore no striae. Like Miller, he reasoned that these stones had been oriented as streamlined bodies in the glacier and had been deposited with but little or no change in orientation. A statistical grouping of long-axis orientations indicates the direction in which the glacier was moving. Later studies at Engebrae and Fondalsbrae glaciers in Norway verified his conclusions. Stones imbedded in debris zones near the end of the glacier show a statistical preference for long-axis orientation parallel with the direction of glacier flow. The majority of stones in the till beside the glacier have similar orientations. Although Richter's conclusions are essentially those of Miller and Bell and appear to be a rediscovery of the earlier knowledge, his conclusions are supported by quantitative, statistical data.

Richter noted that many of the larger till stones (small boulders) at the Norwegian localities tend to an orientation transverse to glacier flow. These were thought to have been oriented originally parallel with the ice-flow direction and to have been shoved during a temporary readvance of the glacier; or to have been rolled beneath the glacier because they projected through the lowermost shear layer (Scherpakete). But the small cobbles in the same deposit are oriented essentially parallel with the ice-flow direction. This, he believed, may have resulted from the action of later meltwater streams which were competent to reorient the smaller stones, leaving the larger ones unmoved.

In describing the Pleistocene glaciation of a part of Yellowstone National Park, Miner (1937) cited the phenomenon of surface boulders with their long axes in parallel orientation, trending diagonally across a valley, but he offered no special explanation.

Krumbein (1938, p. 273; 1939) has investigated statistically the axial orientation of till stones in the western Great Lakes region. He concluded that both the mode and the arithmetic mean of long-axis directions approximate the direction of glacier flow at the locality studied.

The present writer (Holmes, 1938) published a preliminary statement on till-stone orientation in central New York and suggested till fabric as a term denoting the space relations among the component rock and mineral fragments in undisturbed till.

PRESENT PROBLEM

The till in central New York shows the usual preferred arrangement of stones parallel with the direction of glacier flow. In addition, it contains many stones whose long-axis orientation is transverse to the known direction of flow, thus giving two statistically preferred positions of the long axes. At a few localities the transverse orientation predominates. Therefore inference as to the direction of glacier flow, based on mere statistical preference in long-axis positions at those localities, would be wholly unreliable. Richter's published data indicate that transverse orientation exists to some extent in the till of northern Germany, though he did not discuss it, and he apparently regarded it as merely accidental. The hypothesis of reorientation of stones by meltwater streams is inapplicable to the till of central New York. The till in that region has a few sandy or silty lenses but it is prevailingly clayey and lacks evidence of vigorous meltwater action. The stones in transverse position lie closely adjacent both horizontally and vertically to those aligned in other directions, and the enclosing till matrix is essentially uniform at any given locality.

The present work is a search for the factors that have caused some stones to assume the transverse position while others, deposited contemporaneously, are aligned parallel with the direction of glacier flow.

Believing that the results of this study could be applied directly in field practice, conclusions have been based on data readily ascertained and recorded in the field rather than on subsequent laboratory study of materials collected. The nature of pertinent data and the techniques used in their collection and analysis may be unfamiliar to most readers. Hence, brief explanations of these subjects are necessary.

ACKNOWLEDGMENTS

Field study was done during the summers of 1937 and 1938. In the preparation of results, the writer is greatly indebted to Professor Richard Foster Flint of Yale University. Professor Adolph Knopf and Dr. Eleanora Bliss Knopf also have given valuable suggestions and assistance.

This paper is part of a dissertation presented for the degree of Doctor of Philosophy in Yale University.

FIELD DATA AND TECHNIQUE

LOCATION AND CHARACTER OF THE AREA

Investigations were made at 31 localities distributed over the Cazenovia and Tully quadrangles and the northern part of the Pitcher quadrangle, near the margin of the Allegheny Plateau in central New York (Fig. 1). The region is hilly and strongly glaciated, with several prominent valleys leading north to the Ontario Lowland. Altitudes of station localities range from about 600 feet to 1900 feet above sea level and are located so as to include a wide variety of topographic situations. Some are within the belt of Valley Heads end-moraine (Fairchild, 1932) and in the border zone of Valley Heads drift beyond that moraine. Others are located north of the moraine zone, and several lie south of the limits reached by the Valley Heads glacier. Striae and drumloidal forms indicate that the Valley Heads glacier came into the region from the northwest. An earlier Wisconsin glacier, extending far beyond the limits of the area, came from the northeast.

SCHEDULE OF FIELD DATA

General statement.—The kinds of data collected are best explained by reference to the schedule used in the field. They are primarily a record of the form, roundness, dimensions, and attitude of unselected stones as they were removed from their original positions in undisturbed till. In a field classification of till stones, adequate control of the unavoid-

able subjective element is attained by requiring a few major distinctions rather than many minor ones. Because only statistical values are sought, the magnitude of probable error introduced by the subjective factor is not objectionable.

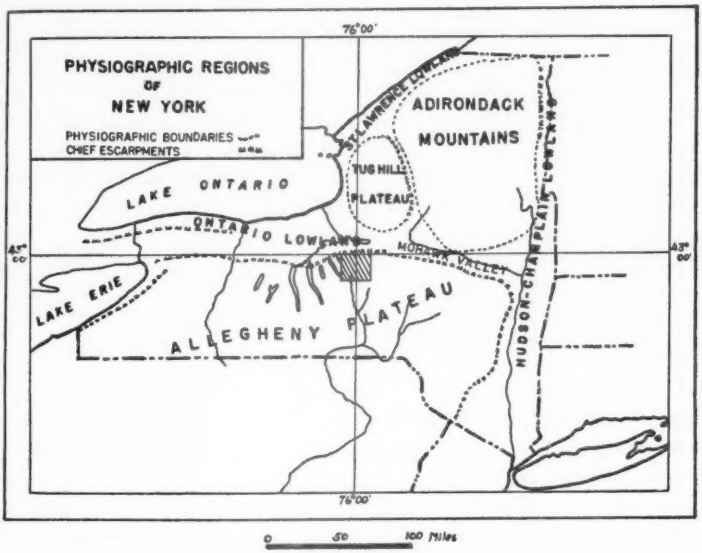

FIGURE 1.—*Map showing location and physiographic relations of area (diagonally lined) in which till-fabric study was made*

Elements of the schedule.—Preliminary field studies evolved the schedule outlined in Table 1, which has been used in this work, although possible improvements were noted as the work progressed. The scheme is adapted to columnar arrangement on single pages of the field notebook.

The long and intermediate axes of a stone (Axis 1 and Axis 2, respectively) are considered to lie at right angles to each other and represent generally the two longer dimensions. Axis 1 lies (ideally) parallel to the more prominent sides if the latter are parallel or bisects the angle between them if they are convergent. The axes need not intersect at their centers. The position of the third or short axis, corresponding to the third dimension of the stone, was not recorded.

The length, width, and thickness of each stone were recorded to the nearest centimeter. Stones less than 0.5 centimeter thick were discarded, except tabular stones whose length and width were 3 centimeters or

more but whose thickness was less than the standard requirement. In such cases the thickness was recorded arbitrarily as one centimeter. This lower limit of size was the only basis for accepting or rejecting stones, though, as a rule, stones longer than 25 centimeters were avoided

TABLE 1.—*Notebook arrangement for recording field observations in till-fabric study as used in field work during 1938*

Direction and dip recorded in 5-degree intervals; size measured to nearest centimeter; form and roundness shown by index numbers and letters. (*See* Plate 1, figure 1.)

No.	Axis 1		Axis 2 Dip	Size	Form	Round-ness	Lithology	Remarks
	Direc-tion	Dip						
1.....	N45E	10NE	5NW	5x4x3	4	d	limestone	Faint striae
2.....	65E	15SW	75SE	2x1x1	3	b	red sandstone	Chipped
3.....	30W	.....	.....	3x2x1	5	c	shale	Striated
4.....	25W	5SE	.....	4x3x1	6	c	shale	

where the number desired from any one station did not exceed 100. The numerical percentage of stones longer than 25 centimeters is small at most localities in the region.

The term form is used in preference to shape, the latter having been defined in terms of sphericity (Wadell, 1932). Sphericity values are believed to be of secondary importance in a field study of glacial stones whose "shapes" are mostly those of the original fragments derived by the breaking-up of larger stones or of the source ledge.

Each stone was classified in one of six major forms (Pl. 1, fig. 1): discoid, ovoid, tabular, wedge-form, rhombohedroid, and varihedroid.[2] Excepting the last two, these terms are in common use in describing pebbles. Rhombohedroids have two sets of parallel or subparallel sides, the ends being either regular or irregular. Varihedroids have a variety of surfaces so related as to render the stone too irregular, unsymmetric, or nodular to be assigned to any of the other forms. Wedge-forms include those whose surfaces converge either laterally or longitudinally. Many have surfaces converging both laterally and longitudinally. The "flat-iron" form described by Von Engeln (1930) and the pentagonal wedge-shaped forms of Wentworth (1936) are included in the longitudinal-wedge group. The terms discoid and ovoid were interpreted rather freely, for

[2] Mr. E. H. Cobb has suggested the term phalangoid for wedge-form; and Professor Carl O. Dunbar of Yale University has suggested heterohedroid as a substitute for varihedroid.

the degree of symmetry required by strict definitions is rare in till stones. Broken discoids and ovoids preserving about half the original forms were found to be numerous and were recorded as separate subgroups.

As explained on a later page, tabular and discoid stones should be grouped together. Their only essential difference is in irregularity of outline, which is more a matter of roundness than of form. Some tabular stones are also related to rhombohedroids but they generally lack the second set of parallel surfaces that characterizes rhombohedroids.

Four degrees of roundness are recognized in this study (Pl. 1, fig. 1). For large pebbles, the approximate equivalents in terms of scales published by Wentworth (1923) and by Tester (1931) are as follows:

Holmes	Wentworth (1923)	Tester (1931)
Roundness "*d*" (sharply angular)	0.00–0.01	0– 18
"*c*" (slightly rounded)	0.02–0.20	19– 45
"*b*" (moderately rounded)	0.21–0.30	46– 75
"*a*" (well rounded)	0.31–1.00	76–100

Stones fractured after a considerable degree of rounding had been attained presented special problems in roundness classification. If only a smoothed side was inherited from a larger stone, or if the previous rounding was slight, it was ignored. Two roundness values were recorded for the half-ovoids and half-discoids. For other stones, a single value was given, based on the relative prevalence of the two degrees of roundness represented.

FIELD PROCEDURE

A mattock was used to prepare a horizontal surface about 2 feet long and one foot wide in an exposure of undisturbed till, from which successive layers could be removed until the desired number of stones was obtained. The stones were uncovered carefully with a prospector's pick. In most instances it was necessary to hold the stone firmly in place until its top and sides were fully exposed. Its long-axis direction was then determined by the use of an orientometer (Fig. 2) in conjunction with a reference rod. The latter is a straight wooden rod driven horizontally, in north-south position, into the bank above the working surface. The orientometer bar was then held horizontally above the exposed stone, and parallel with the reference rod, while the scale beam was rotated until it coincided with the trend of the long axis of the stone. The long-axis direction was then read to the nearest 5-degree interval. Axial dips were ascertained by holding the orientometer so that the bar was horizontal and the plane of the scale beam vertical, and the scale

then rotated until the beam was parallel with the dipping axis of the stone.

Although the statistical results showed that 50 stones generally suffice to indicate the preferred axial orientations, the minimum recorded at any one station was 100 stones, the maximum 800.

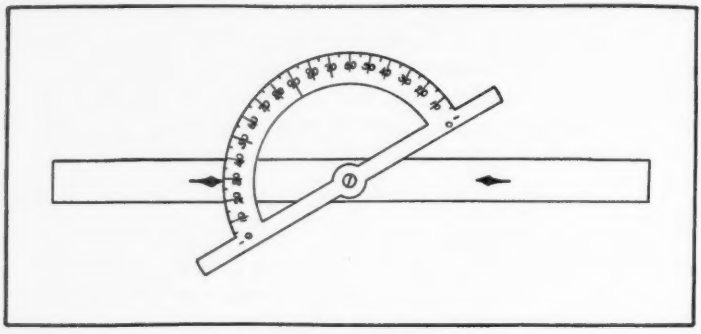

FIGURE 2.—*Orientometer*

Frame or bar is made of hard wood 15 centimeters long. Scale is of metal and is attached to the bar by a screw through the middle of the scale beam. Both margins along the back side of the bar are marked off in centimeters for measuring axial lengths of the stones.

GRAPHIC REPRESENTATION OF TILL FABRIC

General statement.—The "rose" diagram and the contoured stereographic diagram were found the most useful in representing till fabric. The latter diagram is preferred for some purposes, but the former is expressive, easily constructed, and adequate for most requirements. Diagrams were drawn first of the unselected group of stones at each of the 31 stations. Later, selected groups of stones were diagrammed for purposes of comparison.

Rose diagram.—The first step in constructing the rose diagram is to tabulate the long-axis directions (Fig. 3). Then, using a suitable scale, direction lines are drawn for each 5-degree interval, the length of each line representing the number of stones recorded for that direction. The rose diagram expresses details of angular relations that are obscure in the table.

Contoured stereographic diagram.—A contoured stereographic diagram has the recognized advantage of a three-dimensional presentation of any single element and therefore is superior to the rose diagram when distribution of both axial direction and axial dip are to be shown. The

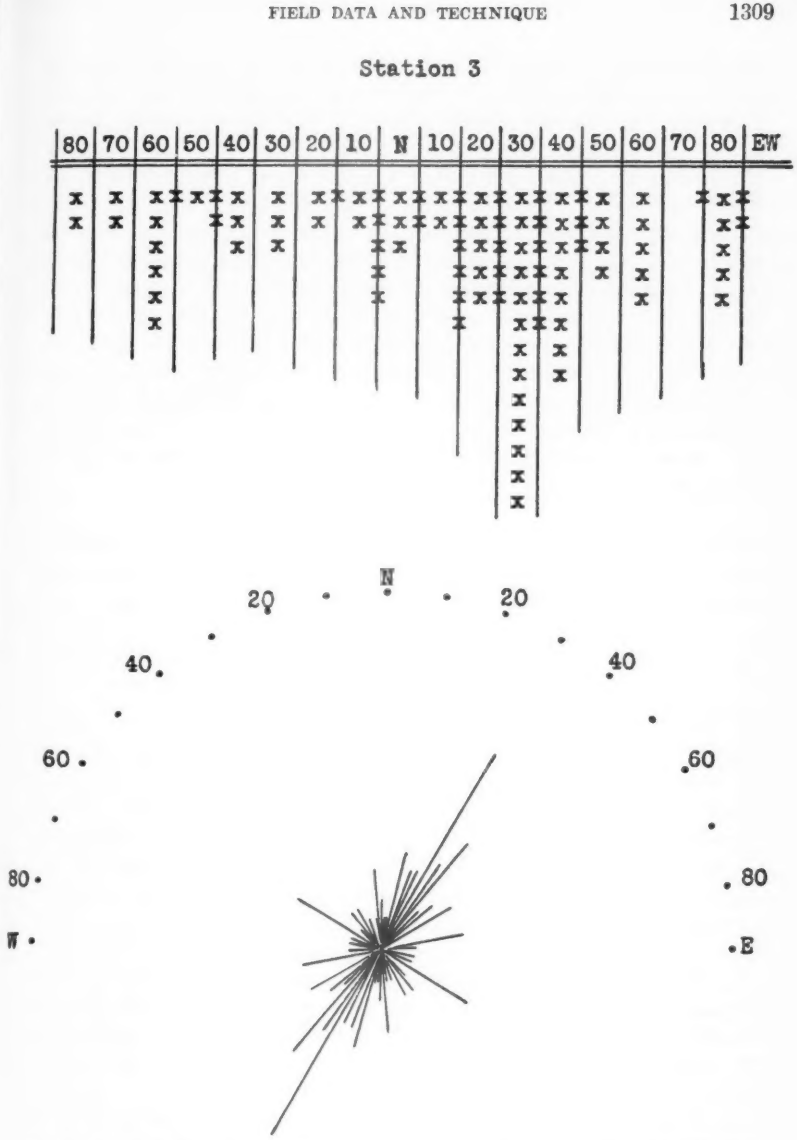

FIGURE 3.—*Distribution chart of long-axis directions, and corresponding rose diagram drawn to a convenient scale*

technique of its construction is fully explained in a recent work by Knopf and Igerson (1938). In brief, the method as here used consists of plotting the long-axis orientations on a Schmidt equal-area projection of a hemisphere 20 centimeters in diameter. This hemisphere is conceived as the inside lower half of a sphere whose limiting great-circle plane is horizontal. With the center of this plane as a reference point, compass directions are marked along the periphery. Dips are recorded from zero degrees at the periphery to 90 degrees at the center, according to the mathematical requirements of the projection. The "percent concentration" of the projected axes or poles at any one point on the projection is the number of poles occurring within a given radius from that point. In this till-fabric study a counting circle having an area equal to 2 per cent of the projection surface has proved satisfactory. Counts are made on a one-centimeter grid over the whole surface of the projection. From the numbers thus derived and spaced, isopleths are drawn at convenient intervals.

STATION PATTERNS

General statement.—A rose diagram of the 100 or more unselected stones from any one station has its own unique features of lengths and angular relations of lines. No two are alike, and it became convenient to refer to each diagram as a fabric pattern. Thus each of the 31 stations has its own peculiar pattern, showing one or more preferred directions of long-axis orientation. Figure 4 shows representative station patterns.

Normal "simple" and "compound" patterns.—On nearly all patterns, the longest line (or one of the longer ones where two or more are present) parallels approximately the direction of glacier flow as indicated by the nearest striae or drumloidal axis and by the topographic environment of the station. Other prominent lines may intersect the main one at angles similar to those of intersecting sets of striae commonly observed on rock ledges. At some localities a pattern derived at a given horizon in a section shows preferred orientations slightly different from those of patterns representing higher or lower zones. At one station, two such zones are separated by a sandy layer about 4 inches thick bearing comparatively few stones. At another station in moderately sandy-silty till, similar zones one or two inches thick are superposed with no perceptible boundaries. Other similar occurrences were noted. A pattern that represents only one horizon is likely to have a single prominent line indicating the direction of glacier flow and is a relatively "simple" pattern. Less prominent lines occur on either side of the main one and are progressively shorter with increasing angular distance in the

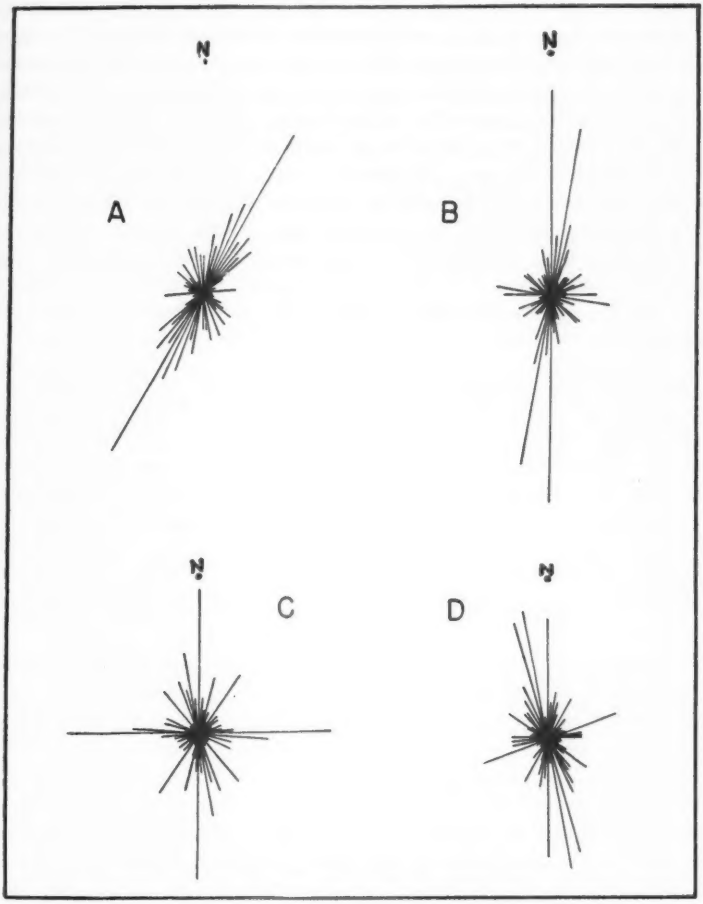

FIGURE 4.—*Typical till-fabric patterns from localities in central New York*
(A) Branch of Mud Creek valley, 1.7 miles south of Lincklaen Center, Pitcher quadrangle.
(B) Shallow through-valley, 0.75 mile northeast of Carpenter Pond, Cazenovia quadrangle.
(C) Limestone Creek valley, 1.8 miles southeast of Oran, Cazenovia quadrangle.
(D) Upland, 3.5 miles northwest of Pompey, Cazenovia quadrangle.

pattern. By contrast, many patterns represent more than one till horizon and therefore present a "compound" appearance. However, both are alike in having their longest lines coincident with the ice-flow direction, a feature normal to the majority of fabric patterns.

Transverse patterns.—A feature common to most station patterns is the presence of one or more prominent lines essentially at right angles to the inferred direction of glacier flow; and in a few patterns such transverse lines are much longer than the others. Thus arises a possible (though unlikely) alternative interpretation that in those places the glacier had flowed in a direction at right angles to that indicated by other available evidence. Moreover, a few of these stations are so situated that glacier flow could be inferred in either of the two right-angle directions indicated by the patterns. These patterns are a departure from the "normal" fabric and are conveniently designated as "transverse" patterns. The diagram in Figure 3 is known to be a transverse pattern and is discussed in detail on a later page as a "problem pattern."

Interpretations.—Superposed fabric patterns in a till section are interpreted as a record of slight shifts in direction of glacier movement while the till was accumulating beneath the glacier by a "plastering-on" process (Fairchild, 1907). Such changes in direction may or may not have been accompanied by intermittent glacier flow. A pressure-melting temperature has been shown to exist at the base of large glaciers (Hubbert, 1937). Hence constant deposition of the basal load—a "plastering-on"—seems probable except under conditions tending to depress the pressure-melting-point isogeotherm into the floor beneath the glacier.

Presumably the factors responsible for orientation of a few till stones transverse to glacier flow at many places are responsible also for the dominantly transverse orientation observed at some localities. These factors must be found in order that patterns showing dominantly parallel or transverse preference may be distinguished. The field facts warrant the conclusion that the preferred directions of long-axis orientation of unselected groups of stones are not infallible independent criteria for determining the direction of glacier flow. Further analysis is necessary.

FABRIC ANALYSIS

GENERAL METHOD

The field studies were planned to include factors believed most likely to show positive correlation with either parallel or transverse alignment of till stones with respect to the direction of glacier flow. The general plan of analysis is to select groups of stones on the basis of their ascertained characters and depositional attitudes and then to compare the orientation patterns of these selected groups. In this way the relative influence of form and degree of roundness on the behavior of stones

as they were being deposited can be evaluated. Facts thus secured provide independent criteria for distinguishing between parallel and transverse orientation, and hence the direction of glacier flow. Also, inferences can be made regarding the mechanics of glacial deposition and the probable nature of the transportational environment.

The depositional attitude of a stone in undisturbed till is the position attained at the final moment of glacial transportation. The following analyses show that characters inherent in the till stones predispose them to certain depositional attitudes, but factors in the transportational environment are equally important in determining the till fabric.

COMPOSITE GROUP

General statement.—A homogeneous body of data suitable for analysis was obtained by combining the data from those stations that satisfied three requirements: (1) The station pattern must indicate, as nearly as possible, a single direction of glacier movement (like a single set of striae). (2) The direction of glacier movement at the station must be known from other evidence. (3) The distribution of long-axis dips, plotted by stereographic projection, must be reasonably symmetrical. On the assumption that the till accumulated progressively or by successive increments, a symmetric distribution of long-axis dips would indicate that the surface of deposition was essentially horizontal and therefore practically coincident with the imaginary reference plane from which dip is measured.

Of the 31 stations, 10 fulfilled these requirements sufficiently to permit combining them into a composite group, giving a total of 1180 stones. The several station patterns were combined by rotating each one until its line representing direction of glacier flow was in the conventional north-south position. Thus all lines in the composite pattern (Fig. 5A) have reference only to the known direction of glacier flow and not to compass directions. The direction from which the ice approached is indicated on this and on all subsequent diagrams (except those of Figures 6 and 23) by a dot above each diagram.

Characteristics of the composite group.—Two significant features of the composite group should be mentioned before discussing selected subgroups. First is the three-dimensional distribution of long-axis positions as shown by the contoured diagrams. Figure 5B shows that the long-axis poles tend to concentrate along two girdles. The more conspicuous girdle occupies the peripheral zone of the diagram and includes both the parallel and the transverse maxima. The other girdle, less well defined, includes the parallel maximum but lies normal to the transverse

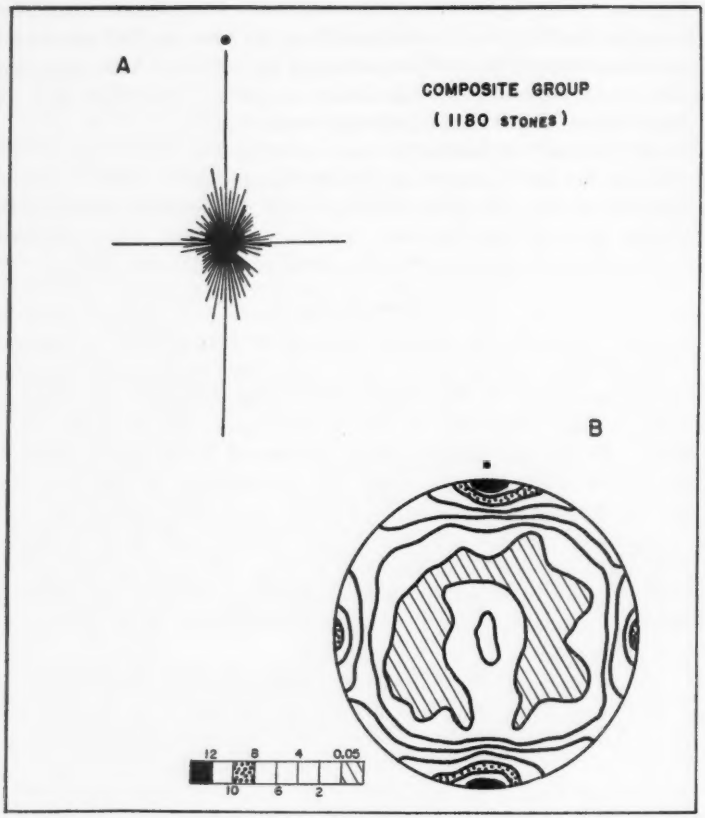

FIGURE 5.—*Rose and contoured diagrams representing combined till-fabric patterns from 10 localities*

Each individual pattern was superposed by placing its recorded ice-flow direction in north-south position. Thus the direction from which the ice approached is indicated by a dot above each composite diagram.

maximum and includes a prominent concentration of steeply dipping poles near the center of the diagram. These facts suggest that the presence of this secondary girdle may be a reliable guide to the direction of ice flow. Inspection of the contoured diagrams for the several stations (Fig. 6) shows that such a girdle is present in some but is indistinct in many, possibly because of an insufficient number of stones represented in each.

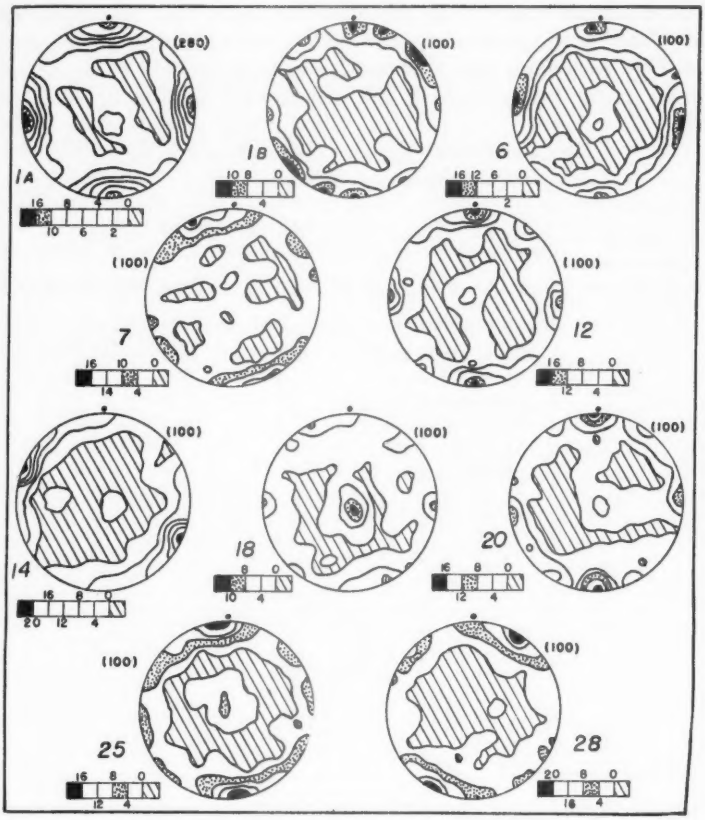

FIGURE 6.—*Contoured till-fabric diagrams of the 10 stations in composite group*

1A, 1B, 6, and so forth, are field station numbers. Number of stones represented in each diagram indicated in parentheses. True north shown by dot above each diagram. Note indistinct percent-concentration bands (girdles) across some of the diagrams, and compare with Figure 5B.

A second important feature of the composite group is the numerical distribution of axial dips through the entire range of dip. Table 2 shows this distribution for both long and intermediate axes. More than one-third of the total number of stones were imbedded with neither axis dipping as much as 10 degrees. With increasing amounts of dip, the numerical frequency declines sharply at about 20 degrees but rises again between 70 and 90 degrees. Table 3 shows how form and roundness

of till stones are related to the distribution of axial dips. In conjunction with Table 4, which gives the per cent of the composite group represented by each form and roundness, roundness appears to be an important factor in determining the depositional position of some till

TABLE 2.—*Numerical frequency in amounts of axial dips*

Composite group (1180 stones).

Dip of intermediate axis (degrees)	Dip of long axis (degrees)																		
	0	5	10	15	20	25	30	35	40	45	50	55	60	65	70	75	80	85	90
0	332	41	73	25	29	10	16	3	11	3	5		6	3	9	5	19	8	1
5	21	17	11	3	1				1	1							1	3	
10	36	14	19	7	4	2	1		1	1	3		1	2	2	2	5		
15	11	4	10	8	4	2	1	1	2	1						4			
20	17	4	12	1	3		2		1				1						
25	6	3	3	1		1	1												
30	10	1	1	1	1		1	1					4						
35	5																		
40	12	2	6	1	1	1			1	1									
45	3									2									
50	13		3	2					3										
55				1		1	1	2											
60	12	2	7	1	1	2	3												
65	5	2	1	2	2	3													
70	12	1	4	4	8														
75	8	4	4	10															
80	29	6	24																
85	37	12																	
90	22																		

stones. However, no conclusions are attempted on the basis of these charts alone.

SUBDIVIDING THE COMPOSITE GROUP

Figure 7 shows the general scheme used in subdividing the composite group. Subdivisions were made, first, on the basis of till-stone characters, and second, according to amounts of axial dips. In general, increasing contrasts in orientation patterns appear with increasing restriction in subgroup selection. The statistical value of groups containing less than 20 stones is best appraised by noting their contribution to the patterns of the next larger subdivisions. To serve as an index to the parallel or transverse positions, the orientation pattern of any subgroup must show appreciable variations from the composite pattern (Fig. 5).

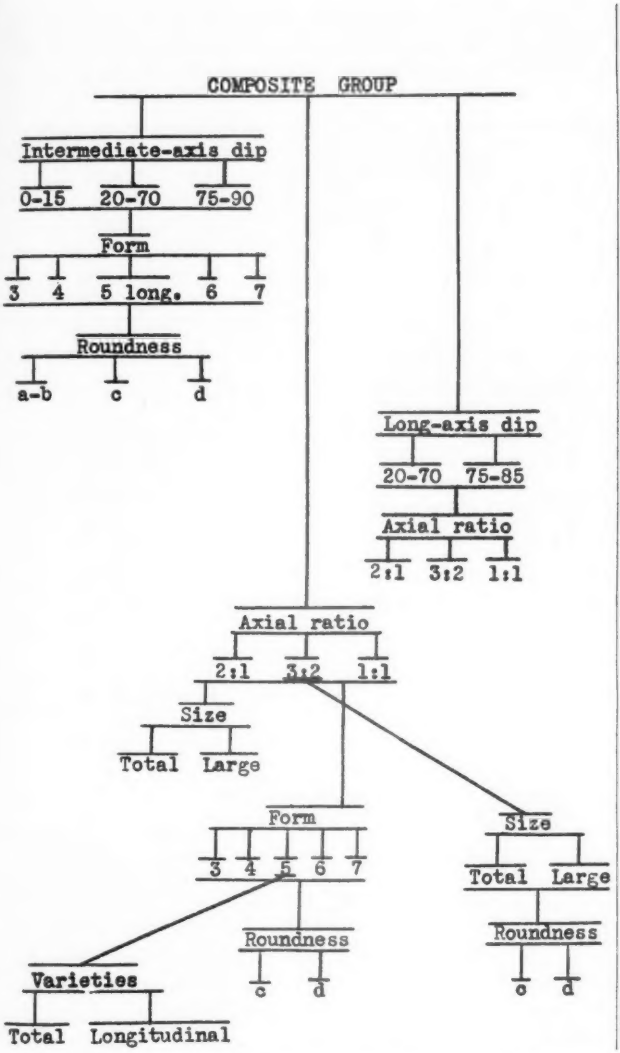

Figure
5
8
9-12
16
17
18-22

(a-d) Degrees of roundness; (3) Ovoid; (4) Rhombohedroid; (5) Wedge-form; (6) Tabular; (7) Varihedroid.

FIGURE 7.—*Diagram showing plan of analyzing composite group*

Index to figures in right-hand column. Figures 13 to 15, not listed, present subdivisions based only on form and roundness.

Comparative percentages in the parallel and transverse positions are of chief importance in evaluating the orientation preference of a selected group of till stones. In addition, percentages within a range of 10

TABLE 3.—*Numerical relation of form and roundness to degrees of axial dips*
Composite group (1180 stones).

Dip of intermediate axis	Form	Dip—long axis												Roundness
		0°—15°				20°—70°				75°—90°				
		a	b	c	d	a	b	c	d	a	b	c	d	
0° to 15°	2	1	3	2			2	1						
	3	3	34	19		1	4	1			4	2		
	4			121	56			27	15			4	3	
	5			125	114			25	21			11	9	
	6			15	21			6	4			1	1	
	7			45	50			8	7			4	5	
	8			1	5								1	
	9			7	11			2	2				3	
20° to 70°	2		1	2										
	3	1	7	1			1							
	4			31	24			12	2					
	5			28	31			11	8					
	6			11	7			3	1					
	7			1	8			3	6					
	8			1	2									
	9			1	3			2						
75° to 90°	2		2											
	3		7	3										
	4			23	17									
	5			33	30									
	6			3	5									
	7			8	15									
	8				1									
	9			5	5									

Forms:	Number:
2—Discoid	14
3—Ovoid	87
4—Rhombohedroid	336
5—Wedge-form	451
6—Tabular	78
7—Varihedroid	162
8—Half-discoid	11
9—Half-ovoid	41
Total:	1180

degrees on both sides of these positions also are significant because they suggest the extent of near-parallel and near-transverse alignment. The orientation diagrams show these values approximately, but in some respects Table 5 affords more accurate comparisons.

Contrary to expectations, subgroups established on the basis of form and roundness show less contrast in orientation preference than appears among subgroups based on axial dip, particularly intermediate-axis dip. The latter subgroups are presented first.

TABLE 4.—*Percent distribution of form and roundness among the 10 stations of the composite group*

Numbers at column headings indicate form; letters indicate roundness.

Station	2	3		4		5		6		7		8	9	No. of stones
		a–b	c	c	d	c	d	c	d	c	d			
1	1	9	7	32	10	16	7	3	2	4	3	1	5	280
1B	0	5	1	12	5	23	27	0	0	7	8	4	8	100
6	3	7	1	21	7	19	23	6	9	1	1	0	2	100
7	1	2	0	20	16	13	27	3	4	3	10	0	1	100
12	1	2	1	11	17	20	28	3	2	2	9	0	4	100
14	0	4	0	12	13	24	18	3	5	12	7	0	2	100
18	1	3	0	14	7	14	22	1	7	11	19	0	1	100
20	2	7	1	13	5	32	9	5	2	11	3	3	7	100
25	2	5	2	8	8	21	26	4	3	4	13	1	3	100
28	0	0	1	16	11	25	20	7	1	7	12	0	0	100
Average	1.2	5.0	2.4	18.3	9.9	19.9	18.7	3.3	3.3	5.9	7.7	0.9	3.5	

Forms: 2—Discoid
3—Ovoid
4—Rhombohedroid
5—Wedge-form
6—Tabular
7—Varihedroid
8—Half-discoid
9—Half-ovoid

INTERMEDIATE-AXIS DIP IN RELATION TO LONG-AXIS ORIENTATION

The tendency of till stones to occur with their intermediate axes dipping either less than 20 degrees or more than 70 degrees makes possible a three-fold subdivision of the composite group on the basis of amounts of intermediate-axis dip. The limits chosen are arbitrary but are designed to restrict the end members to relatively small but equal arcs. Figure 8 shows the rose diagrams of the three groups thus established.

Comparison of these diagrams with that of the composite group shows that intermediate-axis dips not exceeding 15 degrees are statistically associated with parallel orientation; that transverse orientation is likewise associated with steeply dipping intermediate axes; and, similarly,

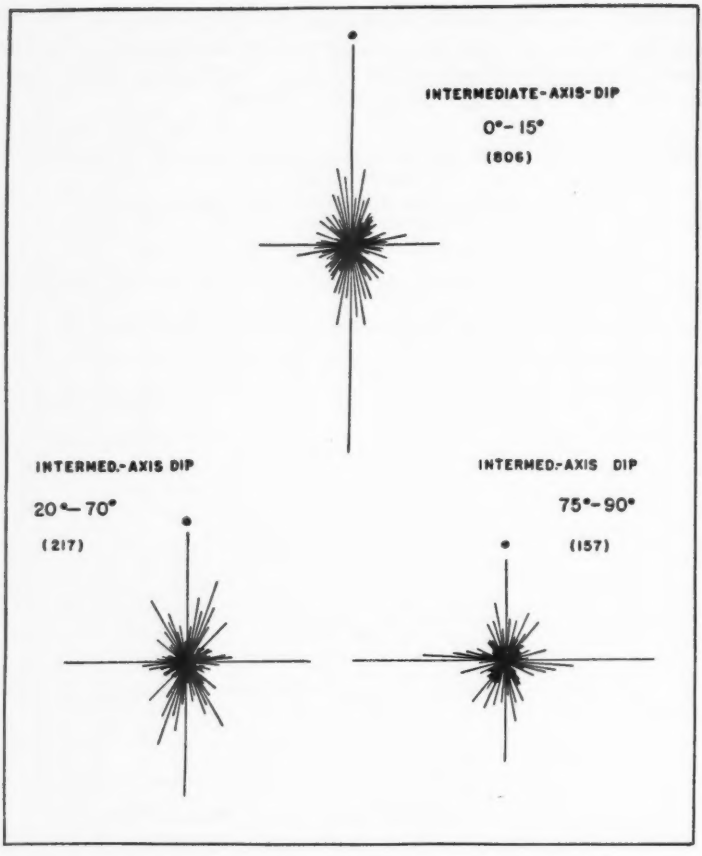

FIGURE 8.—*Orientation patterns of groups based on amounts of dip of intermediate axes*

Numbers of stones represented shown in parentheses. Comparison of the three diagrams shows that steep dip of intermediate axis is correlated with long-axis direction transverse to glacier flow; and that low dip of intermediate axis is correlated with long-axis direction parallel with that direction.

that intermediate-axis dips ranging from 20 to 70 degrees are correlated with diagonal orientation. These facts show that the relative scarcity (per unit range or interval) of long-axis dips ranging from 20 to 70 degrees (shown by Table 2) cannot be attributed to postdepositional compaction of the till. Although compaction would reduce the amount

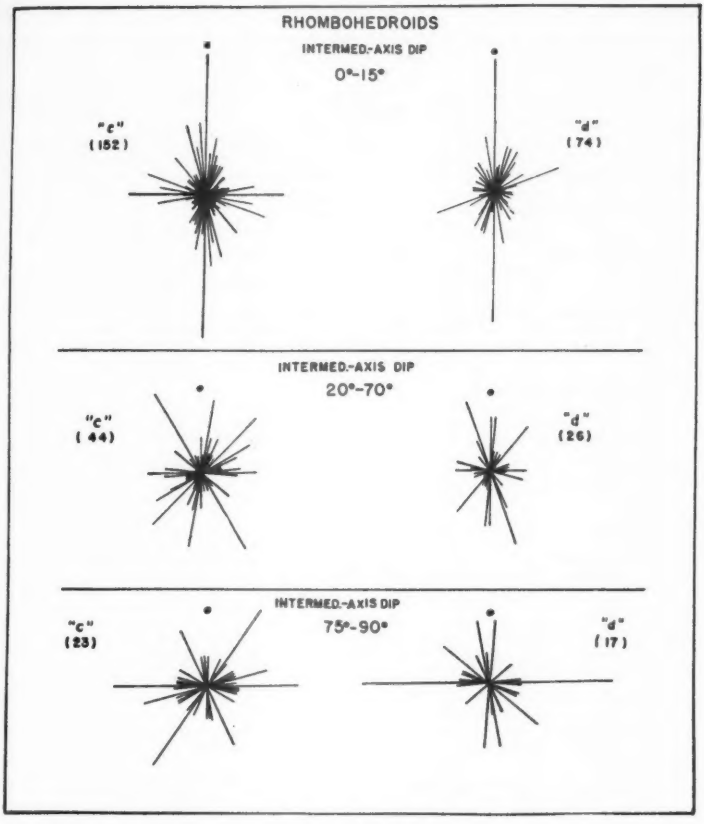

FIGURE 9.—*Orientation patterns of rhombohedroids from three groups shown in Figure 8, further classified as to roundness*

Slightly greater contrast appears among the "*d*" patterns than among the "*c*" patterns; but the contrast averages somewhat greater among these axial-ratio groups than is shown by those of Figure 8.

of tilt of imbedded stones and would be most effective within the 20 to 70-degree range, it would not at the same time exercise a selective change in long-axis direction. Hence compactness of the till with its imbedded stones is believed to be an original depositional character and seems to be logically related to gradual accumulation ("plastering-on") beneath the weight of the moving glacier; it implies a well-defined contact between the ice and its till floor.

TABLE 5.—*Tabular comparison of till-stone groups as represented by rose diagrams*

Percentages included within a range of 10 degrees from the parallel and transverse positions are given, as well as the percentages occurring precisely in those positions.

[Fig.]	Pattern	No. of stones	Per cent		Per cent (within 10°)	
			Parallel	Transverse	Parallel	Transverse
5	Composite group, total	1180	11.4	7.0	27.0	17.5
8	Intermediate-axis dip 0°–15°	806	12.9	5.7	30.0	16.0
	Intermediate-axis dip 20°–70°	217	9.5	8.5	21.7	18.5
	Intermediate-axis dip 75°–90°	157	7.6	11.5	18.5	24.2
9	Rhombohedroids					
	Intermediate-axis dip 0°–15°, "c"	152	11.9	6.6	29.8	18.6
	Intermediate-axis dip 0°–15°, "d"	74	16.7	1.4	29.8	8.3
	Intermediate-axis dip 20°–70°, "c"	44	4.6	6.8	18.2	18.2
	Intermediate-axis dip 20°–70°, "d"	26	11.6	7.4	26.9	15.4
	Intermediate-axis dip 75°–90°, "c"	23	4.5	13.6	13.6	22.8
	Intermediate-axis dip 75°–90°, "d"	17	0.0	23.5	29.4	35.2
10	Longitudinal-wedge-forms					
	Intermediate-axis dip 0°–15°, "c"	91	13.2	3.3	37.4	11.0
	Intermediate-axis dip 0°–15°, "d"	104	16.3	2.9	32.7	14.4
	Intermediate-axis dip 20°–70°, "c"	27	11.1	7.4	18.5	18.5
	Intermediate-axis dip 20°–70°, "d"	31	13.9	6.5	19.4	19.4
	Intermediate-axis dip 75°–90°, "c"	22	13.6	4.5	13.6	9.1
	Intermediate-axis dip 75°–90°, "d"	20	5.0	5.0	10.0	23.0
11	Ovoids					
	Intermediate-axis dip 0°–15°, "a"–"b"	46	21.8	6.5	41.3	15.2
	Intermediate-axis dip 0°–15°, "c"	22	9.1	9.1	18.3	27.3
	Intermediate-axis dip 20°–70°, total	9	0.0	11.1	22.2	44.4
	Intermediate-axis dip 75°–90°, total	10	0.0	10.0	0.0	30.0
11	Varihedroids					
	Intermediate-axis dip 0°–15°, "c"	57	5.3	8.8	22.8	21.1
	Intermediate-axis dip 0°–15°, "d"	63	9.2	3.2	25.8	11.3
	Total	162	7.4	8.3	22.2	17.9
12	Tabular					
	Intermediate-axis dip 0°–15°, "c"	22	13.6	9.1	45.5	18.2
	Intermediate-axis dip 0°–15°, "d"	26	19.3	3.8	38.5	7.7
	Intermediate-axis dip 20°–°70, "c"	14	21.4	0.0	35.7	7.3
	Intermediate-axis dip 20°–70°, "d"	8	12.5	12.5	25.0	25.0
	Intermediate-axis dip 75°–90°, "c"–"d"	8	12.5	25.0	12.5	25.0
	Discoid, total	14	29.0	15.0	36.7	35.7
13	Rhombohedroids, "c"	219	9.7	7.5	25.8	19.2
	Rhombohedroids, "d"	117	14.0	6.0	29.0	13.7
13	Wedge-forms, total, "c"	230	12.4	8.9	28.5	16.9
	Wedge-forms, total, "d"	221	10.8	4.1	26.7	17.1
14	Ovoids, "a"–"b"	59	17.0	6.7	30.2	20.2
	Ovoids, "c"	28	10.7	5.1	17.9	28.6
14	Varihedroids, "c"	72	7.0	9.7	22.5	19.4
	Varihedroids, "d"	90	7.8	7.8	22.2	16.8
15	Tabular, "c"	39	12.9	5.3	38.5	12.9
	Tabular, "d"	39	18.0	7.7	33.3	15.4
15	Half-ovoids, "c"	17	11.8	5.9	29.4	17.6
	Half-ovoids, "d"	24	16.7	4.2	25.0	12.5
	Half-discoids, total	11	0.0	9.1	9.1	9.1
	Discoids, total	14	29.0	15.0	35.7	35.7
16	Long-axis dip 20°–70°					
	Axial ratio *2:1*	37	14.5	14.5	29.7	14.5
	Axial ratio *3:2*	90	12.2	4.4	25.5	8.8
	Axial ratio *1:1*	47	12.8	2.1	27.7	10.6

TABLE 5.—*Tabular comparison of till-stone groups as represented by rose diagrams*—Continued

[Fig.]	Pattern	No. of stones	Per cent		Per cent (within 10°)	
			Parallel	Transverse	Parallel	Transverse
	Long-axis dip 75°–85°					
	Axial ratio *2:1*	6	0.0	33.3	0.0	33.3
	Axial ratio *3:2*	27	18.5	0.0	55.1	7.1
	Axial ratio *1:1*	14	14.2	0.0	42.8	14.2
17	Axial ratio *2:1*, total	260	11.1	8.1	28.1	13.5
	Axial ratio *2:1*, minimum length 5 cm.	29	17.3	13.8	27.6	20.7
	Axial ratio *3:2*, total	639	11.0	6.8	26.8	19.1
	Axial ratio *3:2*, minimum length 5 cm.	184	10.8	5.0	33.2	16.3
	Axial ratio *1:1*, total	281	12.5	6.5	26.6	17.8
	Axial ratio *1:1*, minimum length 5 cm.	24	4.2	16.7	12.5	33.3
18	**Rhombohedroids**					
	Axial ratio *2:1*, "c"	42	14.6	9.7	24.4	14.6
	Axial ratio *2:1*, "d"	40	7.5	12.5	32.5	17.5
	Axial ratio *3:2*, total, "c"	129	10.0	5.4	29.4	17.0
	Axial ratio *3:2*, total, "d"	47	19.2	2.1	29.8	17.1
	Axial ratio *3:2*, minimum length 5 cm "c"	55	11.1	3.7	40.8	18.5
	Axial ratio *3:2*, minimum length 5 cm "d"	16	12.5	0.0	31.3	18.8
	Axial ratio *1:1*, "c"	48	4.2	10.6	17.0	27.7
	Axial ratio *1:1*, "d"	30	10.0	3.3	23.3	3.3
19	**Wedge-forms, total**					
	Axial ratio *2:1*, "c"	60	13.2	6.7	30.0	13.3
	Axial ratio *2:1*, "d"	46	8.7	6.5	30.5	8.4
	Axial ratio *3:2*, total, "c"	111	9.0	9.0	24.3	16.2
	Axial ratio *3:2*, total, "d"	121	9.9	4.1	23.1	22.3
	Axial ratio *3:2*, minimum length 5 cm "c"	25	4.0	12.0	16.0	16.0
	Axial ratio *3:2*, minimum length 5 cm "d"	31	12.9	3.2	25.9	16.1
	Axial ratio *1:1*, "c"	59	18.7	10.2	33.9	22.0
	Axial ratio *1:1*, "d"	54	14.9	1.8	31.5	14.8
20	**Longitudinal-wedge-forms**					
	Axial ratio *2:1*, "c"	36	8.3	2.8	33.3	8.3
	Axial ratio *2:1*, "d"	30	6.7	6.7	33.3	10.0
	Axial ratio *3:2*, "c"	72	13.9	2.8	29.1	8.3
	Axial ratio *3:2*, "d"	82	12.2	3.7	26.6	19.5
	Axial ratio *1:1*, "c"	34	14.7	8.7	32.3	26.5
	Axial ratio *1:1*, "d"	42	19.0	0.0	26.2	16.7
21	**Ovoids**					
	Axial ratio *2:1*, total	11	27.2	0.0	27.2	9.1
	Axial ratio *3:2*, total, "a"–"b"	39	15.4	2.6	33.3	23.1
	Axial ratio *3:2*, total, "c"	21	9.5	14.3	28.6	28.6
	Axial ratio *3:2*, minimum length 5 cm.	13	23.1	0.0	46.2	15.4
	Axial ratio *1:1*, "b"–"c"	16	6.3	18.7	12.6	25.0
22	**Varihedroids**					
	Axial ratio *2:1*, "c"	20	14.3	14.3	19.0	14.3
	Axial ratio *2:1*, "d"	21	4.8	4.8	33.3	14.3
	Axial ratio *3:2*, "c"	35	5.7	8.6	28.5	25.7
	Axial ratio *3:2*, "d"	52	7.7	11.5	21.1	17.3
	Axial ratio *1:1*, "c"	16	0.0	6.2	12.5	12.5
	Axial ratio *1:1*, "d"	17	11.8	0.0	11.8	5.8
22	**Tabular**					
	Axial ratio *2:1*, total	6	0.0	0.0	33.3	0.0
	Axial ratio *3:2*, "c"	33	15.2	6.1	30.3	18.2
	Axial ratio *3:2*, "d"	28	14.3	10.7	25.0	21.4
	Axial ratio *1:1*, "a"–"c"	18	22.2	11.1	50.0	22.2
	Axial ratio *1:1*, "d"	10	30.0	0.0	50.0	0.0

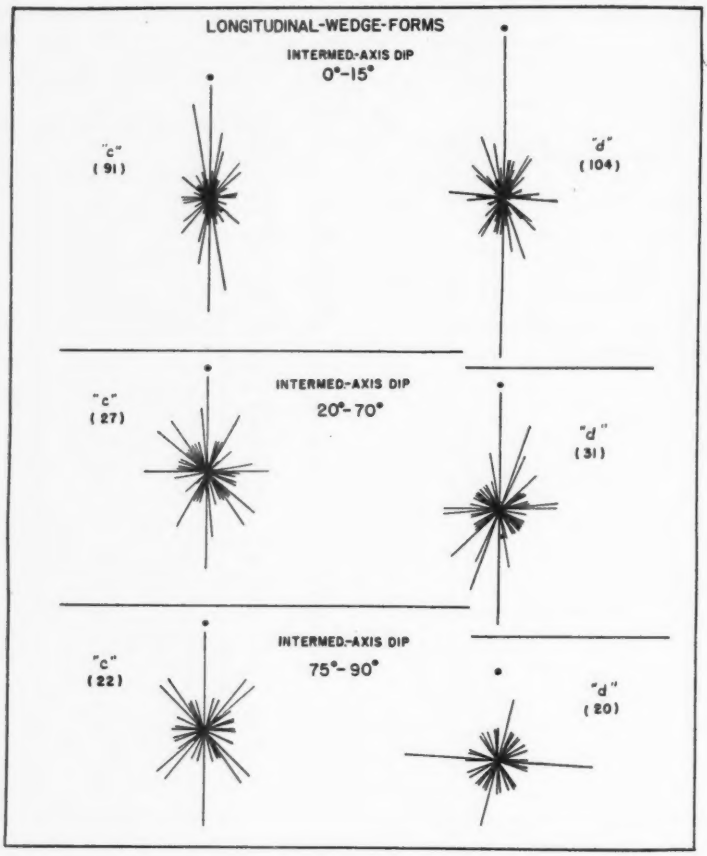

FIGURE 10.—*Orientation patterns of longitudinal-wedge-form stones from three groups shown in Figure 8*

Transversely aligned stones tend to an approximate rather than a precise alignment in that position. Patterns of lateral-wedge-forms resemble those of rhombohedroids and are not shown.

The diagrams in Figure 8 show that the attitude of the plane that includes the two longer axes of a till stone is as important as the direction of the long axis. Neither rose nor contoured diagram can show both simultaneously, but the three rose diagrams together constitute an approximation within the limits of the dip-interval selected.

The next step in analysis is to determine how the different forms and degrees of roundness of till stones influenced long-axis direction in each of the three groups just considered.

COMBINED INFLUENCE OF FORM, ROUNDNESS, AND DIP OF INTERMEDIATE AXIS ON LONG-AXIS ORIENTATION

General statement.—Figures 9 to 12 show the three groups (based on amounts of intermediate-axis dip) further subdivided on the combined bases of form and roundness, except those subgroups that are too small to warrant plotting. These diagrams present some of the greatest contrasts brought out by any basis of subdivision.

Among rhombohedroids of "*d*" roundness, not one in the 75 to 90-degree group is oriented in parallel position; while in the 0 to 15-degree group only one occupies the transverse position (Fig. 9). Therefore, on the basis of intermediate-axis dip, these two groups of sharply angular rhombohedroids are good statistical indicators of the transverse and parallel positions, respectively. Rhombohedroids of "*c*" roundness show the same tendencies but to a smaller extent. Both "*c*" and "*d*" rhombohedroids show a statistical avoidance of the parallel and transverse positions when their intermediate axes dip within the 20 to 70-degree range.

Much less contrast appears among the longitudinal-wedge-forms [3] (Fig. 10), though in the 75 to 90-degree group, roundness "*d*" presents one of the strongest transverse patterns of any wedge-form group. However, slight rounding ("*c*") established a parallel-orientation preference. Excepting this small group of sharply angular stones, wedge-forms show a consistent preference for parallel or near-parallel orientation somewhat stronger than that of other forms.

Low dip of intermediate axis is characteristic of a large majority of ovoids. (Fig. 11.) Those of roundness "*a*" and "*b*" show also an extraordinary preference for parallel orientation, while those less perfectly streamlined ("*c*") have a decided preference for transverse or near-transverse alignment. Ovoids of "*c*" roundness approach varihedroids "*c*" in general relative proportions and appear to resemble them in depositional habit.

Further appreciation of the contrasts in orientation preference already noted is gained by comparing the diagrams in Figures 9 to 12 with those of groups based only on form and roundness. Accordingly Figures

[3] Longitudinal-wedge-forms include all wedge-form stones except those with surfaces converging only laterally. The latter have been found to be more closely related to rhombohedroids and are not included here.

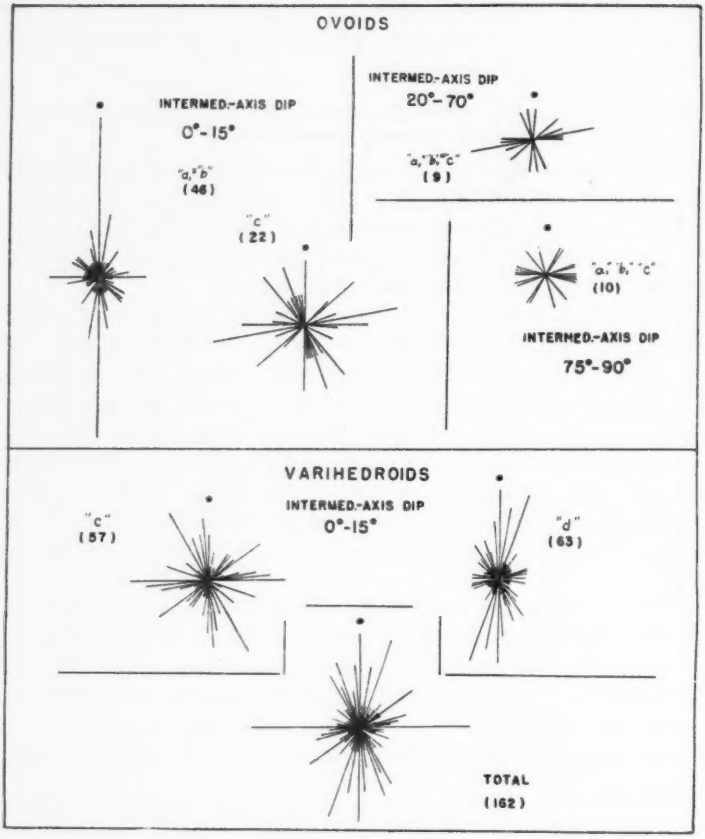

FIGURE 11.—*Orientation patterns of ovoids and varihedroids from the three groups shown in Figure 8*

In the 0-15-degree group, roundness has a greater influence on long-axis direction of ovoids than on that of other till-stone forms. Varihedroids show the least consistency in orientation of any of the till-stone forms. However, rounding transforms many varihedroids into general ovoid forms, which shows in similarity of the corresponding "*c*" patterns.

13 to 15 have been prepared. The long-axis orientation of discoids appears to be closely related to that of tabular stones (Fig. 15). Therefore the discoids are combined with the later stones of "*c*" roundness in the contoured diagram as Tabular "*a*," "*b*," and "*c*." . The group of least rounded ("*c*") ovoids is too small to be plotted stereographically by

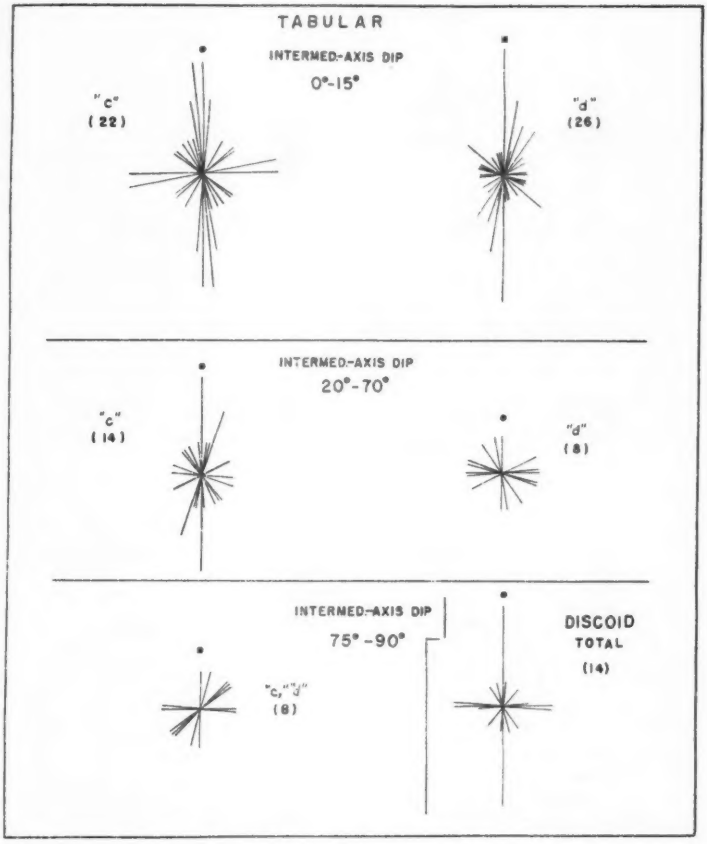

FIGURE 12.—*Orientation patterns of tabular stones from three groups shown in Figure 8*

Composite group contains only 14 discoids (mostly "*a*" and "*b*" roundness), and the discoid pattern is included here to show its similarity to that of the "*c*" tabular stones.

itself, but its general characters can be evaluated by comparing the two contoured diagrams of the ovoid group (Fig. 14).

SYMMETRY OF FORM

The diagrams show that varihedroid stones (Fig. 14) are comparatively lacking in orientation preference, apparently because of their corresponding lack of symmetry. By contrast, ovoids of "*a*" and "*b*" roundness have the strongest tendency to assume or to maintain an orientation

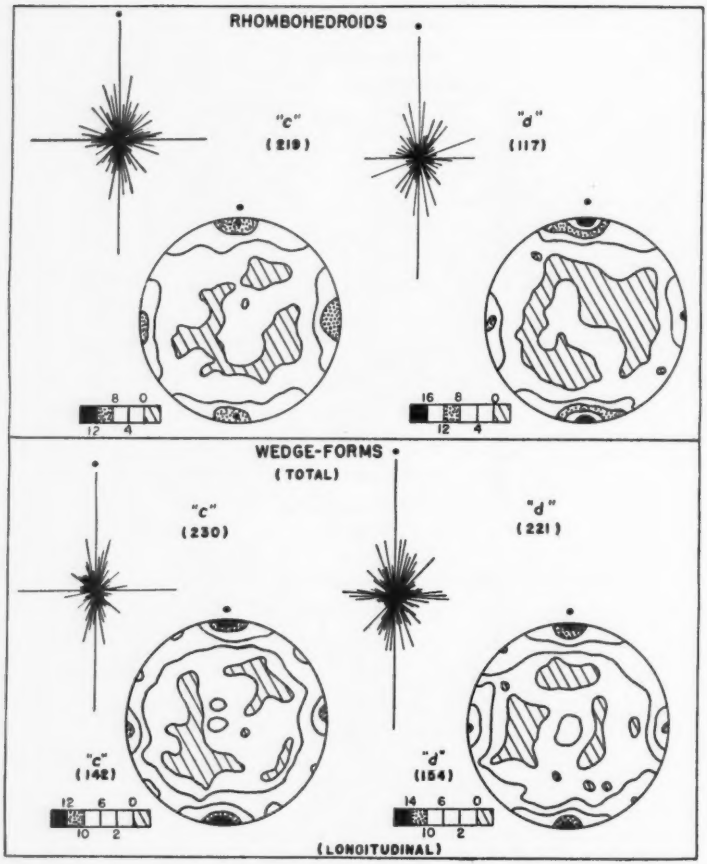

FIGURE 13.—*Orientation diagrams of rhombohedroids and wedge-forms from composite group shown in Figure 5 subdivided according to roundness*

Rose diagrams of wedge-forms include both lateral and longitudinal types. Lateral-wedge-forms are omitted from contoured diagrams.

in harmony with the direction of glacier movement. These ovoids are the most perfectly streamlined of all till-stone forms. Rhombohedroids are comparable to varihedroids in roundness, but their degree of symmetry is more nearly that of ovoids, and their degree of orientation preference is pronounced. Hence it is concluded that symmetry of form conduces to strength and uniformity of depositional orientation.

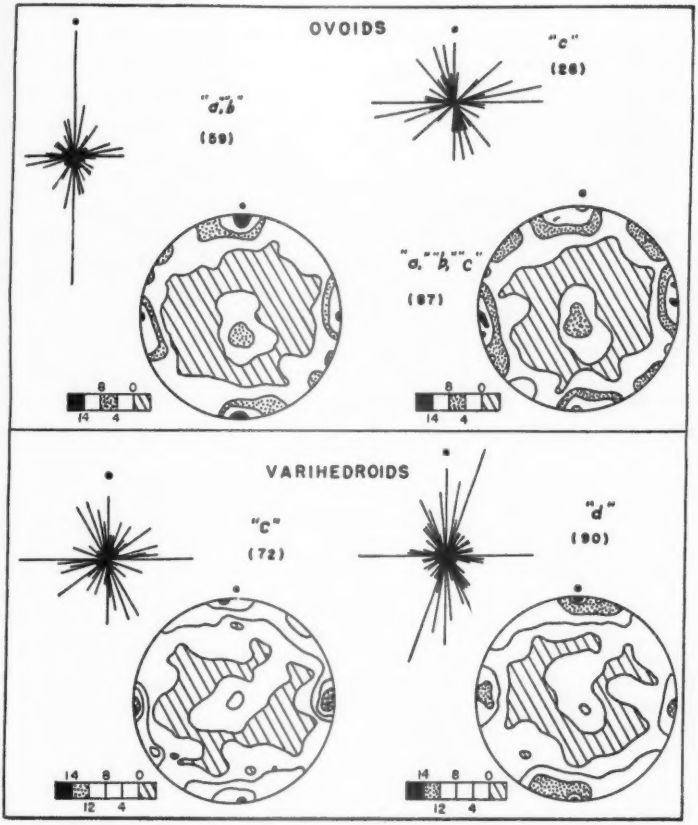

FIGURE 14.—*Orientation diagrams of ovoids and varihedroids from composite group shown in Figure 5, subdivided according to roundness*

Contoured diagram of ovoids at right includes total number of these forms in composite group. Note similarity of patterns of "*c*" groups.

CONTROL BY PROMINENT SURFACES OF TILL STONES

The trend of striae on till stones is significant inasmuch as such striae record the relative direction of differential movement of the stone and its ice matrix. Striae on rhombohedroids generally parallel the long axis. Because that axis parallels the sides of the stone, the relative importance of surfaces and of long-axis location in controlling the movement of a rhombohedroid in transit cannot be evaluated. But on wedge-form stones the angular relation between sides and long axis permits such

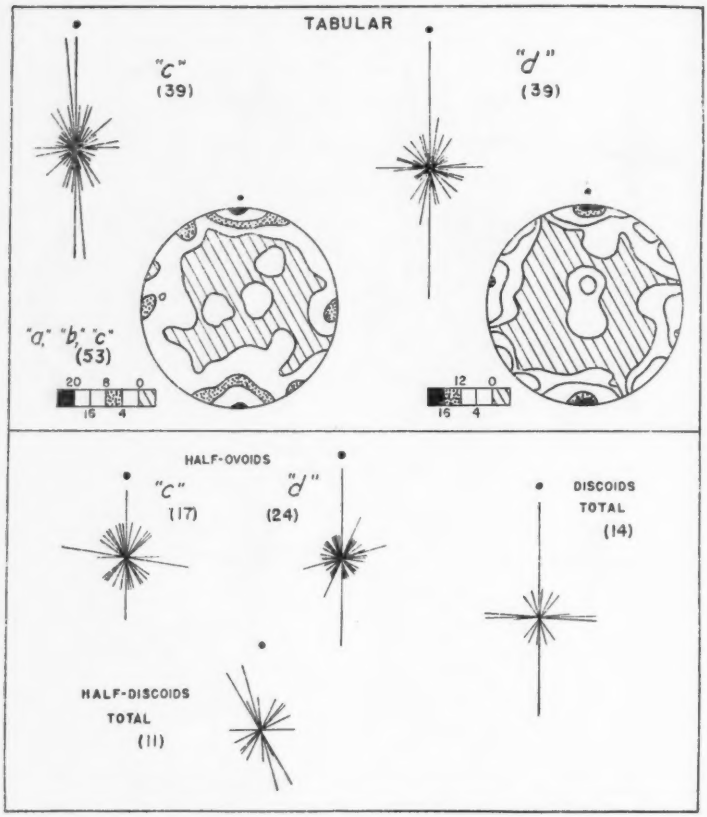

FIGURE 15.—*Orientation diagrams of tabular stones, discoids, half-ovoids, and half-discoids from composite group shown in Figure 5*

Tabular stones and half-ovoids subdivided according to roundness. Half-ovoids and half-discoids are included here for completeness. Diagram of all discoids is shown for comparison with those of tabular stones. The close similarity is apparent.

evaluation. The photograph (Pl. 1, fig. 2) shows that striae on wedge-forms have a strong tendency to parallel the edges of surfaces on which they occur and consequently are parallel to the adjacent surfaces bounded by these same edges. Scarcely any strong striae parallel the long axis of such stones. Photographs of wedge-shaped stones published by Wentworth (1936) show the same tendency. This means that, in the differential movement of stone and matrix whereby the striae originated,

the stone was so oriented that adjacent sides intersecting at a given edge (or possibly two edges and three surfaces) were parallel to the direction of flow at any one time. The position of the stone shifted occasionally as control of orientation passed to other surfaces and edges. Movement in this case was by gliding rather than by rotation and evidently was due to ice pressure upon the rear surface because of greater velocity above the horizon of the base of the stone. The long axis of the stone would seldom be well aligned with the direction of movement unless the pointed end of the wedge were directed toward the ice thrust (rear) and the diverging sides adjusted to a streamlined position.

Thus wedge-form stones show that prominent surfaces exercise a strong control in orientation, but further discussion of the probable transportational behavior of wedge-forms is best deferred to a later page. Rhombohedroids seem to illustrate best the problem of orientation by gliding.

ORIENTATION BY GLIDING

Orientation controlled by prominent surfaces is assumed to be much more effective with rhombohedroids than with wedge-forms because all the four principal surfaces are (ideally) parallel to each other. As these surfaces are likewise parallel to the long axis, rhombohedroids are considered best for studying orientation control in gliding.

Figure 9 shows that rhombohedroids whose intermediate axes dip less than 20 degrees lie predominantly parallel to the direction of glacier flow. The contoured diagrams show that the long axes of most stones thus aligned are likewise essentially horizontal. This is indicated also to some extent by the chart (Table 3). These facts suggest the following interpretation: Where till is accumulating beneath the moving glacier, a stone in contact with the till floor is retarded by friction which may become great enough to overcome the forward force of ice thrust. A rhombohedroid sliding thus in parallel orientation and with the two longer axes horizontal presents the least possible cross-sectional area against ice thrust. Its third (shortest) dimension affords a minimum projection into the faster-moving ice above the floor, and a maximum basal surface is tending to adhere to the deposited till. Such an orientation is imposed by sliding (Sander, 1934) and thus may easily become the permanent depositional attitude unless the stone is shoved by another in transit before it becomes buried in the accumulating till. The great number of apparently random orientations suggests that such shoving is common, but the statistical preference of flat-lying rhombohedroids for parallel orientation is interpreted as the position inherited from their transportational orientation.

ORIENTATION BY ROTATION

The fact that the sides of rhombohedroid forms are parallel to the long axis makes these stones especially adapted to forward movement by rotation. Observed phenomena and experimental analogy suggest that they do roll forward when carried above the base of the ice.

In a recent paper, Demorest (1938) has interpreted the conditions of flow at the base of a small, recently vanished glacier in Glacier National Park. Striae on the irregular rock floor show that the ice had moved by laminar flow comparable to the movement of a viscous liquid, apparently in conformity to the laws of fluid mechanics. An experiment by Taylor (1923) illustrates the kind of movement postulated for stones carried in such a medium.

Taylor studied the motion of small oblate and prolate spheroids immersed in water glass of high viscosity. The containing vessel was cylindrical and mounted on a base capable of rotation on a vertical axis. A uniform shearing motion was imparted by a large stationary rod fastened vertically in axial position within the container. The aluminum bodies rotated in response to the shearing motion. When they had reached stable orientations, the prolate spheroids were rotating about their long axes, the latter being in a vertical position. The oblate spheroids rotated about equatorial diameters in similar orientation. Therefore, by analogy, a rhombohedroid carried above the base of the ice would roll forward on its long axis, with that axis horizontal and transverse to the direction of movement. Ovoids likewise would adjust readily to that orientation. Tabular stones would correspond to the oblate spheroids and may be expected to rotate similarly. Most striae on convex surfaces of till stones are believed to be a record of such rotation in a plastic abrasive medium.

Deposition from the rotational orientation postulated for rhombohedroids would account satisfactorily for long-axis direction transverse to that of glacier flow. This would be possible if movement ceased, as in stagnation, but would fail to explain the association of steeply dipping intermediate axes that is statistically characteristic of the rhombohedroids thus oriented. In theory, continued forward movement should be easiest and most likely when the intermediate axis has reached a dip of about 70 degrees and least likely when continued rotation has brought that axis again to a near-horizontal position. Cessation of glacier flow when the rotating stones were in edgewise position is a statistically improbable coincidence. Moreover, these stones were imbedded in the same compact till matrix with those in parallel alignment, for which all evidence indicates gradual accumulation beneath

the moving glacier. Inasmuch as rotation may conceivably take place at, or in slight contact with, the till floor, the following interpretation of deposition from rotational orientation is offered.

The rhombohedroid stone was rotating in transverse orientation along the till floor of the glacier. With each rotation the intermediate and short axes rose alternately past the vertical, and the intermediate axis, being the longer of the two, required a larger vertical range. Other stones carried slightly higher (and therefore moving slightly faster) entered this zone, intercepted the rising edge of the rhombohedroid, and by continued forward movement thrust the rhombohedroid edgewise into the till floor. When the intermediate axis was nearly vertical (or just past the vertical) the impinging stone was free to move past, leaving the rhombodedroid permanently imbedded.

This hypothesis is equally applicable to ovoids. Inasmuch as ovoids lack the parallel edges and prominent flat surfaces characteristic of rhombohedroids, a greater tendency to axial wabbling may be expected with consequently less accurate transverse alignment when imbedded. The orientation diagrams of ovoids confirm this expectation (Figs. 11, 14).

In contrast with rhombohedroids, wedge-forms are not well adapted to continued rolling in transverse orientation. On rotation, the pointed end would advance less rapidly than the other, and one or two complete rotations would suffice to bring the stone back into parallel alignment. Hence wedge-form stones receive more consistent abrasion by sliding than by rotation, which seems to explain the common observation that wedge-forms are the most likely to bear noticeable striae (Wentworth, 1936). The orientation diagrams indicate that sharply angular ("*d*") wedge-forms are equal in persistent parallel orientation to those of "*c*" roundness. As field observation has shown that the "*d*" wedge-forms are produced from larger stones by crushing, the wedge form is the cause of the presence of striae rather than the result of "controlled" glacial abrasion as interpreted by Von Engeln (1930). Demonstrable abrasion facets are rare, and the majority of those observed in central New York seem best explained as boulder-pavement phenomena.

Thrusting of stones into the till floor in the manner suggested offers an explanation for the numerous diagonal orientations shown conspicuously by many rhombohedroid and wedge-form patterns (Figs. 9, 10, 18, 19, 20). A wedge-form stone, sliding forward with the point foremost and with one of the convergent sides parallel to the direction of transport, may change from gliding to rotation. The initial rotation

would bring the stone up on the other (diagonally oriented) convergent side, with the long axis in a diagonal direction. If during this movement superjacent debris prevented the rear edge from rising far, the stone would be forced down into the till beneath, with the long axis still in diagonal orientation. Likewise, a rhombohedroid in parallel orientation may be first turned until its diagonal dimension is parallel to the direction of gliding (like a double wedge). If rotation is then induced, the initial movement would take place by rolling on one of the diagonally placed sides. Superjacent debris might then intercept the rising edge and press the rhombohedroid into the till floor in permanent diagonal orientation.

The hypothesis of permanent deposition from a rotational orientation not only explains adequately the associated steep dips of intermediate axes but also suggests an explanation for the concentration of steeply dipping long-axis orientations shown by many of the contoured diagrams. Richter (1933) noted similar near-vertical orientations of long axes in northern Germany and attributed them to local vertical currents in the glacier. That hypothesis is inapplicable to the till of central New York because such stones are distributed closely among those of prevailingly horizontal positions for which all evidence indicates gradual accumulation of the till beneath the moving glacier. But the possibility of rotation on an axis other than the longest, with deposition by thrust into the till floor, seems worthy of analysis.

A stone moving forward in rotational orientation (transverse) may assume a parallel orientation on coming in contact with the till floor if forward movement continues. Should the stone then be raised slightly above the floor, as by overrunning a gently sloping obstacle, rotation about the intermediate axis could be induced, and the stone might then be thrust into the till by pressure from superjacent debris which would intercept the rising rear end. When the stone had reached a near-vertical position, the impinging stone would be free to slide past.

To test this hypothesis of rotation on an axis other than the longest, Figure 16 was prepared. Figure 16A represents the preferred orientation of the intermediate axes of stones whose long axes were dipping 20 degrees or more. A long-axis dip exceeding 15 degrees (at any of the 10 stations selected for the composite group) probably means that the stone came to its final position from some movement other than the normal one imposed by sliding. Figure 16A shows a reasonably strong percent-concentration in horizontal transverse position, indicating that many of the stones could have been rotating on their intermediate axes, with the latter essentially horizontal, when their transportational move-

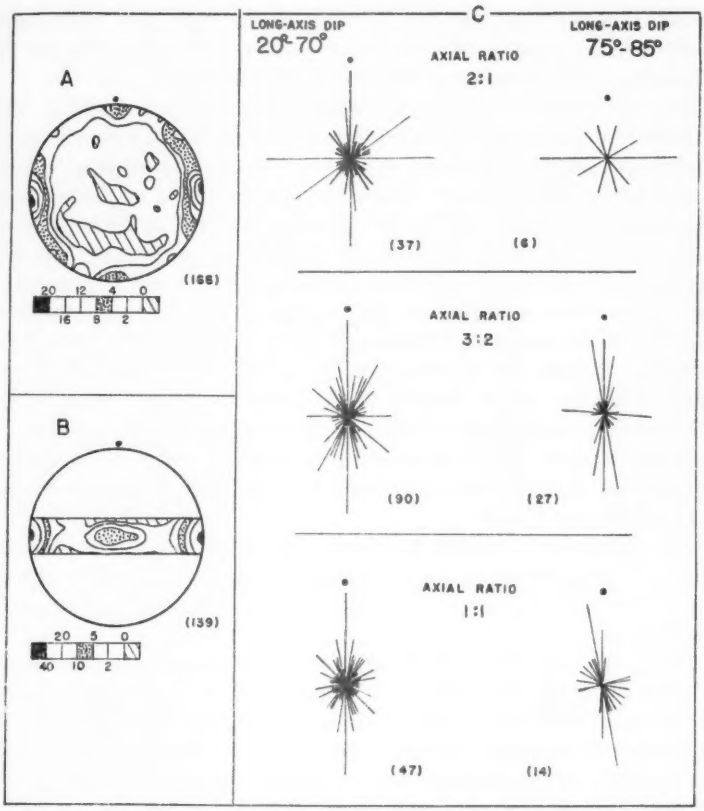

FIGURE 16.—*Orientation diagrams illustrating possible deposition from rotation on one of the two shorter axes*

(A) Diagram showing orientation of intermediate axes of stones whose long-axis dips were 20 degrees or more. The majority of these intermediate axes were in horizontal transverse orientation, which is the normal position of rotation.

(B) Orientation diagram of intermediate axes of stones whose long-axis directions fell within a range of 10 degrees from the direction of glacier flow. Data from four representative stations are shown.

(C) Subgroups based on amounts of long-axis dip and on relative axial lengths. These diagrams should be compared with those in Figure 17, in which these subgroups are included.

ment ceased. Figure 16B is a similar diagram combining four representative stations and showing the percent-concentration of intermediate-axis poles of all stones whose long-axis directions fell within a range of 10 degrees from the direction of ice movement. The diagram in-

dicates that the great majority of these intermediate axes were essentially horizontal and could have served as axes of rotation. Most of the remaining intermediate axes were essentially vertical.

The near-vertical axes are especially significant in connection with Figures 8 to 12 because the shortest axes of those stones were occupying the position normally preferred by the intermediate axes (that is, in horizontal transverse position). Hence the long axes also must have been essentially horizontal. Of the 139 stones represented in Figure 16B, 21 have intermediate-axis dips of 70 degrees or more and comprise the near-vertical group. Of these, seven have a difference of less than one centimeter in the lengths of the two shorter axes, and only five have a difference of more than 2 centimeters in this respect. Therefore many instances of steep dip of intermediate axis are to be explained by the small difference in the lengths of the two shorter axes whereby the shortest may occupy temporarily the position usually assumed by the intermediate axis. Such an explanation probably accounts for most of the parallel-oriented stones represented in the strongly transverse patterns of Figures 8 to 12.

The conclusion indicated by Figure 16A, B is that till stones may rotate to some extent on their intermediate axes, and possibly also on their short axes if the latter are of nearly the same length as the intermediate axes.

Figure 16C and Figure 17 afford additional evidence in support of this conclusion by showing how relative elongation of form may influence depositional orientation. Striae on elongate stones (especially rhombohedroids) are generally parallel to the long axis. However, nearly equidimensional rhombohedroids characteristically acquire intersecting sets of striae arranged parallel to the length (side) and width (end) of the stone. These facts suggest that, where transportational conditions favor sliding, an elongate stone adjusts readily to parallel orientation, while a subequidimensional stone is subequally stable when sliding in either parallel or transverse orientation, presumably controlled by the nearly equal side- and end-surfaces. Likewise, should the mode of forward movement change from sliding to rotation, elongate stones would shift to transverse orientation more readily than would those of subequidimensional proportions.

Figure 17 shows the composite group subdivided on the basis of relative axial proportions (axial ratio) and of size. Stones whose long axes are two or more times the length of the intermediate axes are classed as elongate and for convenience are referred to by the symbol *2:1*. Those

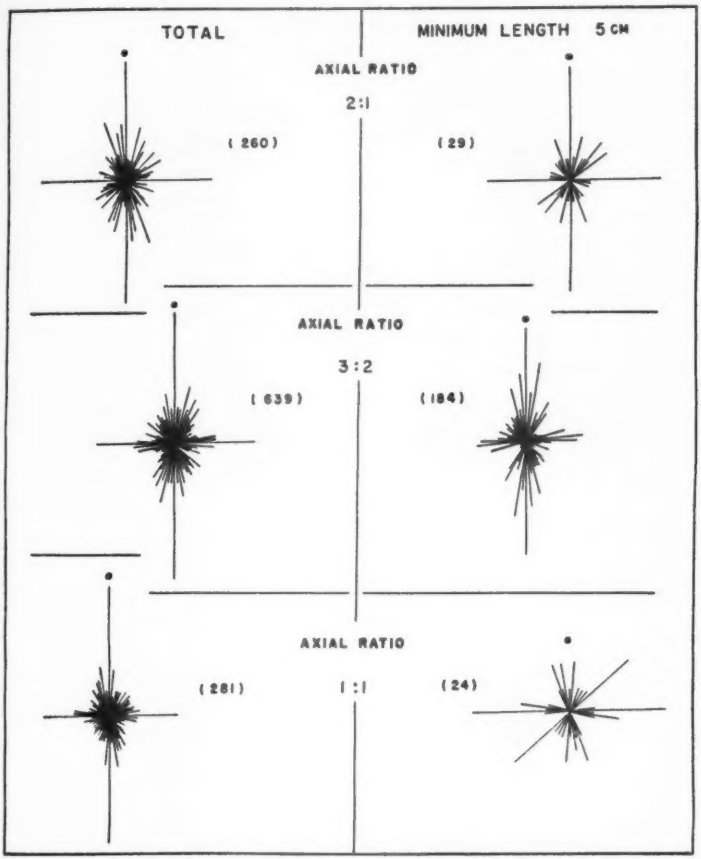

FIGURE 17.—*Orientation diagrams showing influence of axial ratio and relative size*

The three diagrams at the left show composite group subdivided according to relative axial lengths. Stones having long axes 5 cm. or more in length are shown in corresponding diagrams at the right. Note in what ways the larger stones contribute to the total subgroups, especially those of *2:1* and *1:1* axial ratio.

whose axial lengths differ by less than one centimeter (or as near as 7:6) comprise the subequidimensional group, represented by the symbol *1:1*. All other stones are of intermediate axial ratio and are designated by the symbol *3:2*. Except the large *1:1* stones, which have strong transverse preference, the diagrams reveal a slight, though gradual, increase

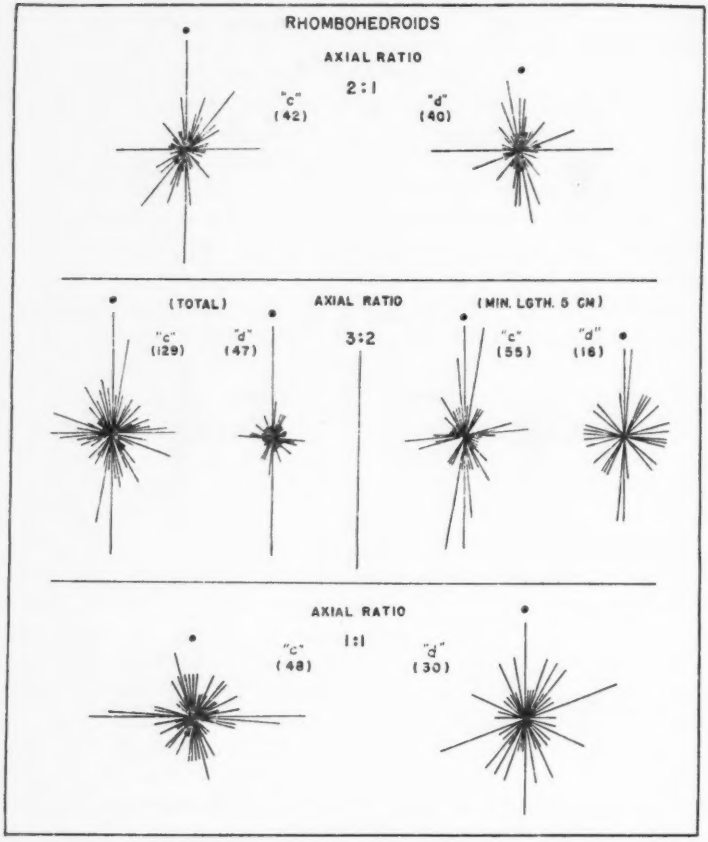

FIGURE 18.—*Orientation diagrams of rhombohedroids from the subgroups (total) shown in Figure 17*

Large stones of *3:2* axial ratio are plotted separately. Influence of roundness on orientation preference is especially marked in stones of *2:1* and *1:1* axial ratio.

in preference for parallel orientation with decreasing difference in axial lengths. From these groups, the stones (213 total) with long-axis dip exceeding 15 degrees are represented in Figure 16C, grouped according to axial ratio and amounts of long-axis dip. The tendency to increased preference for parallel orientation with decreasing difference in axial lengths is appreciably stronger than that shown by the corresponding groups in Figure 17.

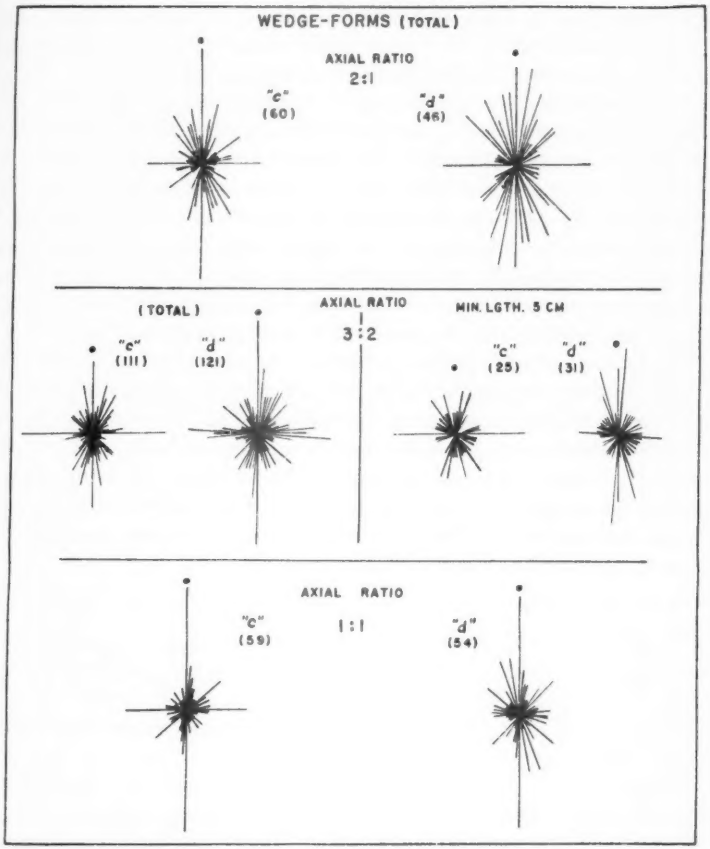

FIGURE 19.—*Orientation diagrams of wedge-form stones from the subgroups (total) shown in Figure 17*

Large stones of *3:2* axial ratio are plotted separately. Among these subgroups, rounding appears to have contrasting effects on orientation preference.

The fact that the *1:1* stones in Figure 16C show a notably strong preference for parallel position, while the amounts of long-axis dip suggest deposition from rotation on the intermediate axis, indicates the probability that the rotational movement began after parallel orientation had been acquired by sliding. Assuming the same transportational history for the other groups in Figure 16C the greatest proportionate loss of parallel orientation before deposition occurred appears in the *2:1* group,

while the *3:2* group shows an intermediate degree of change. The relative influence of form in this connection is suggested by reference to the contoured diagrams in Figures 13 to 15. Diagrams of wedge-form stones show a comparative absence of a girdle normal to the horizontal transverse position, showing that these stones do not lodge readily from rotation on the intermediate axis. This supports the conclusion that any rotation of wedge-forms is more likely to bring them again into parallel orientation. By contrast, the comparatively strong girdle in the ovoid diagrams seems to reflect the relative ease with which a parallel-oriented ovoid may begin rotation on its intermediate axis. If partially imbedded in vertical position, its rounded form would be less subject to dislodging than would the more angular stones under similar conditions.

Summarizing, the following conclusions seem warranted: (1) Stones in glacial transportation may gain and lose parallel orientation at least once, and probably many times, before their final deposition. (2) A parallel-oriented stone in sliding movement may be induced to begin rotation. Normally the initial rotation begins about the intermediate axis, but the long axis soon swings into transverse position and becomes the axis of rotation. (3) The more elongate a stone is, the more readily the long axis assumes transverse orientation.

AXIAL RATIO, FORM, AND ROUNDNESS IN RELATION TO LONG-AXIS DIRECTION

Further analysis of the three groups shown in Figure 17 is significant in showing how axial ratio and roundness may influence preferred depositional orientation of the several till-stone forms. These preferences are illustrated by Figures 18 to 22; though for the small form-groups only the axial-ratio diagrams are given.

Sharply angular ("*d*") rhombohedroids of *1:1* and *3:2* axial ratio (Fig. 18) show a much stronger preference for parallel orientation than is shown by the corresponding groups of "*c*" roundness; but this difference in preference is reversed in stones whose axial ratio is *2:1*. Hence the orientation stability afforded by prominent flat surfaces (discussed previously) evidently does not outweigh the liability of sharp edges and corners of an elongate stone to impinge on obstacles instead of sliding past them. This interpretation supports the belief that elongate stones are the more responsive to conditions tending to change their transportational orientation. On a *1:1* stone, a given amount of rounding destroys a greater percentage of flat "bearing surface" than it would on an elongate stone of equal cross-sectional area, thereby permitting easier rotation and consequently greater chance of deposition from the rotational orientation. The same amount of rounding on an elongate stone would

create only a sled-runner effect, enabling the stone to slide past obstructions.

Another possible factor in the orientation stability of angular stones may be suggested hypothetically by assuming a rhombohedroid oriented by sliding. Its forward velocity is less than that of the ice passing over it. Forward movement of the ice necessitates plastic adjustment at both front and rear, with accompanying pressure-melting and regelation. Adjustment at the front end may take place within a conical zone immediately ahead of the stone. Flow of ice across the top and off the distal end establishes a pressure gradient upward on that end, with the lowest pressure at the upper leading edge of the stone. Hence that edge is the principal locus of regelation. The more sharply angular the edge is, the more concentrated is the zone of regelation. Ice may freeze to that part of the stone, and the general effect of lowered pressure along the upper edge should make the ice in that zone a little firmer than that beneath it. Thus the firm ice may serve as a horizontally asymmetric prow tending to elevate the front end of the stone and to prevent rotation. With slight rounding of the edge, regelation would be no longer sharply concentrated, and the stabilizing effect would be lost.

To apply this principle to the sharply angular *1:1* wedge-forms, it seems necessary to assume that the pointed end faced the ice thrust, with regelation along the broad end of the wedge. Ice thrust against the divergent wedge surfaces might result in more plastic adjustment and less forward movement of the stone, but the long axis would be in close alignment with the direction of transport. The rose diagram (Fig. 20) shows that such alignment is realized to an exceptional degree. If this hypothesis of regelation along a sharp edge is even slightly operative in the manner suggested, it may be the determining factor in holding many angular stones in parallel alignment while finer debris gathers about them, whereas less angular stones would be rolled forward and assume a transverse orientation.

The strongly transverse patterns of the *1:1* ovoids illustrate the tendency of subequidimensional stones devoid of flat surfaces to be deposited from a rotational orientation (Fig. 21). In contrast, ovoids of *2:1* axial ratio show equal persistence in parallel orientation, presumably due to the absence of projecting edges and corners which might otherwise catch on obstructions or on faster-moving debris and thereby cause the stone to turn. Hence these ovoids evidently have a greater resistance to loss of parallel alignment than is characteristic of the *2:1* group as a whole. Ovoids of *3:2* axial ratio show depositional tendencies intermediate between those of the other axial ratio groups.

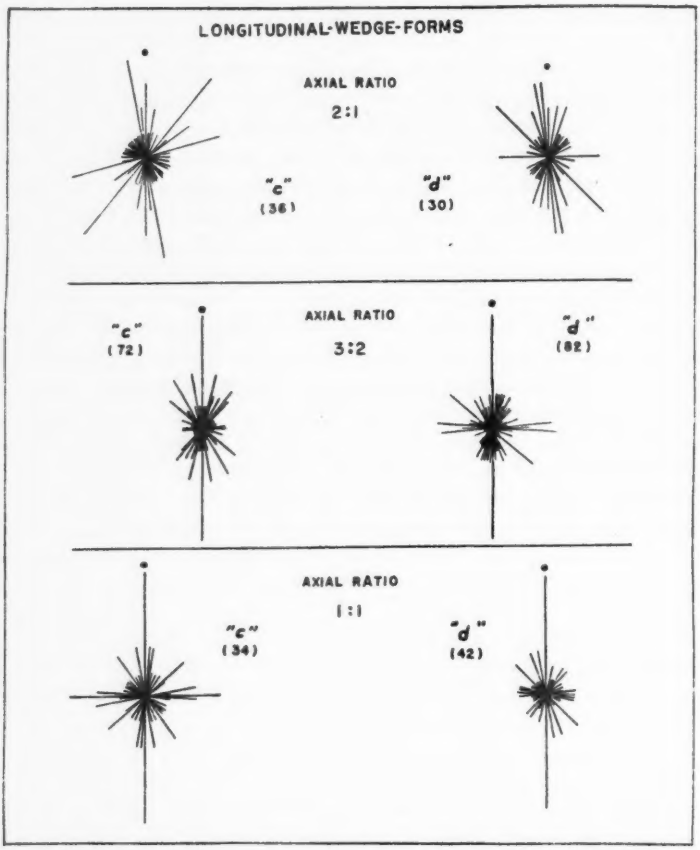

FIGURE 20.—*Orientation diagrams of wedge-form stones, omitting those with surfaces converging only laterally*

Comparison with Figure 19 (in which these stones are included) shows marked differences in the *2:1* and *3:2* groups and only slight contrast in the *1:1* groups.

The depositional behavior of tabular stones deserves brief mention. Although as a group they have a fairly strong tendency to deposition in parallel alignment, they are less likely to be imbedded in horizontal position than other forms are. (*See* Table 3.) Stones with neither long nor intermediate axis dipping more than 15 degrees include 64 per cent of all ovoids; 57 per cent of all varihedroids; 53 per cent of all rhombohedroids and wedge-forms; and only 42 per cent of all tabular stones

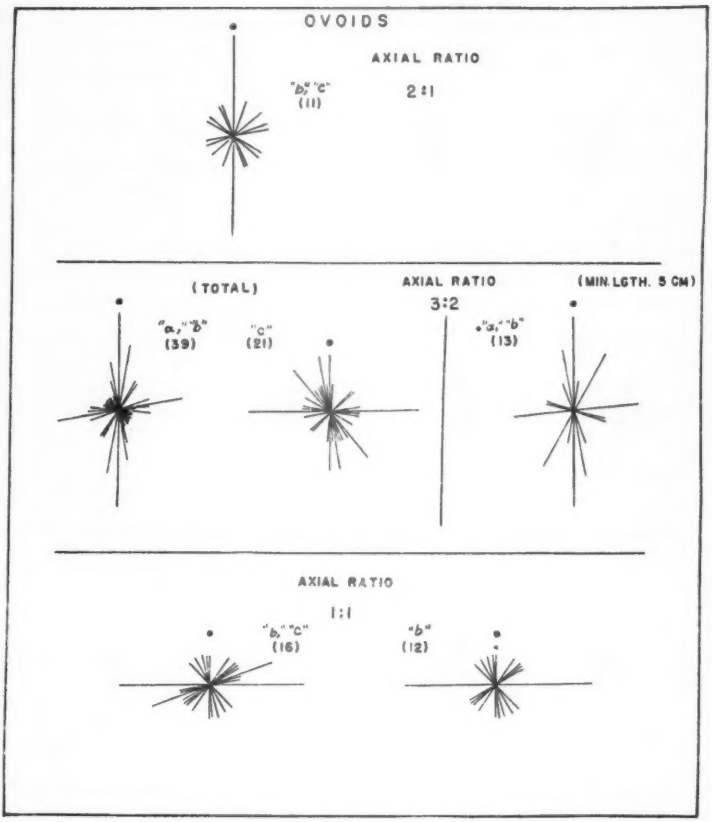

FIGURE 21.—*Orientation diagrams of ovoids from the subgroups (total) shown in Figure 17*

Note marked contrast between the *2:1* and *1:1* groups and the influence of roundness in the *3:2* groups.

(including discoids). These percentages have qualitative significance, for the order of their arrangement is that of the relative thickness of representative stones, that is, length of the shortest axis compared with the lengths of the other two.

Interpreting from Taylor's experiment (described previously), a tabular stone carried above the floor of the glacier would tend to rotate on its long axis. The difference in vertical range required by the two shorter axes at each rotation is greater for tabular stones than for any

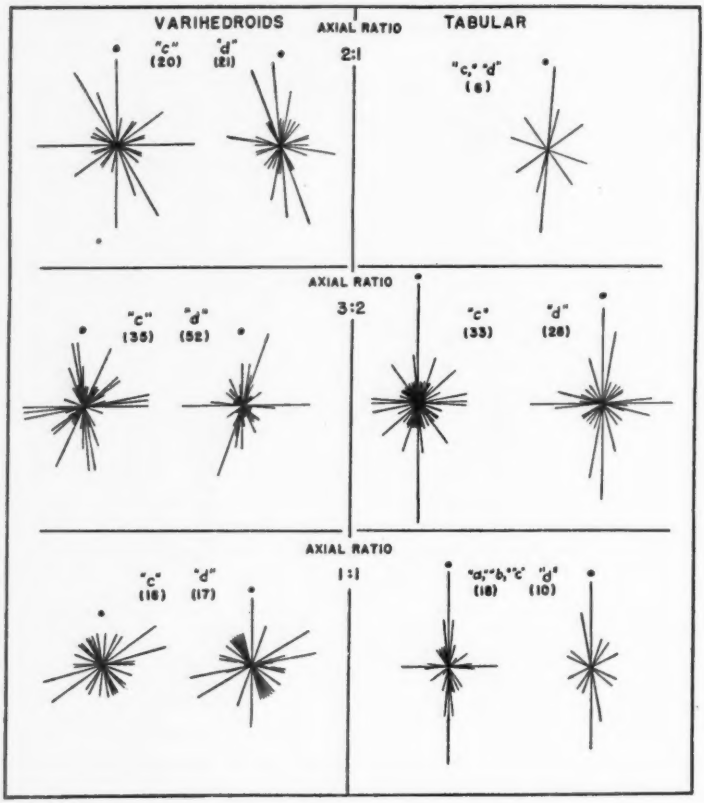

FIGURE 22.—*Orientation diagrams of varihedroids and tabular stones from subgroups (total) shown in Figure 17*

Note characteristic absence of strong alignment in both parallel and transverse positions in most of the varihedroid patterns.

other form, and therefore the longer one (intermediate) would be more likely to encounter superjacent debris in rising past the vertical than would the corresponding axis of any other form. In the debris-laden base of the glacier, such interference may greatly impede rotation. However, the tendency to steeper axial dips suggests deposition by thrust into the till floor, and the preference for long-axis direction parallel to the direction of ice flow then implies at least partial rotation on the intermediate axis. These considerations support the inference

that conditions preventing free rotation tend to compel orientation by sliding and constitute optimum conditions for thrust into the till floor if rotation begins.

CRITERIA FOR RECOGNITION OF ORIENTATION POSITIONS

The foregoing analyses show that till stones differ in statistical preference of depositional position according to their varying inherent characters, modified by the varying conditions of glacial transportation. The variety of groups studied permits the designation of stones whose statistical chance of deposition either parallel with the direction of glacier flow or transverse thereto is somewhat greater than average. Many groups seem to offer possibilities, but the following are considered the best guide forms.

Parallel position: (R. indicates roundness.)

1. Subequidimensional stones R. "*d*," with long axes less than 5 centimeters, except varihedroids.
2. All subequidimensional rhombohedroids R. "*d*."
3. All subequidimensional longitudinal wedge-forms R. "*d*," with intermediate-axis dip not steeper than 70 degrees.
4. The following forms with intermediate-axis dips not steeper than 15 degrees:
   a. Longitudinal wedge-forms R. "*c*" and "*d*," except those of *3:2* axial ratio R. "*c*."
   b. Rhombohedroids R. "*d*."
   c. Ovoids R. "*a*" and "*b*," except subequidimensional.
   d. Tabular R. "*c*" and "*d*."

Transverse position:

1. Subequidimensional rhombohedroids R. "*c*."
2. All subequidimensional stones with long axes at least 5 centimeters in length, except rhombohedroids R. "*d*:" and wedge-forms.
3. Ovoids R. "*c*."
4. Rhombohedroids R. "*c*" and "*d*" with intermediate-axis dips of 75 degrees or more, except subequidimensional R. "*d*."

This list of guide forms contains some duplication. For instance, a subequidimensional longitudinal wedge-form stone, roundness "*d*," may be included in three groups if its length is less than 5 centimeters and if its intermediate axis dips no more than 15 degrees. The statistical value of such stones is probably greater than that of stones qualifying for only one group because they combine a greater number of properties that give orientation stability.

The limitations of these guide forms must be clearly recognized. Their orientation preference is only statistical and relative. Many of them are not abundant in an unselected group, and by the law of chance the first few encountered may possibly show any orientation except their statistically preferred one. Table 5 shows that a frequency of 20 per cent in parallel position, or of 15 per cent in transverse position, is attained in only a few groups.

Pattern habit of the several forms must also be considered. For example, wedge-form stones of "*c*" roundness are likely to be oriented in near-parallel rather than precisely parallel position. Tabular stones of "*c*" roundness show a similar tendency. The value of ovoids of roundness "*c*" as transverse indicators lies largely in their tendency to near-transverse orientation.

The *1:1* rhombohedroids illustrate a critical problem in strong orientation preferences because rounding from "*d*" to "*c*" reverses their preference. The reversal probably requires a certain range in the rounding process, and therefore stones in this transition stage are probably worthless as guides. A variation of the problem appears in groups selected on the basis of intermediate-axis dips (Figs. 9-12). The intervening dip-interval is 50 degrees (20 to 70), and the groups seem well separated. But they really grade into each other as the two shorter axes of the stones approach each other in length. The control exercised by axial-length difference disappears if the two axes are equal.

As mentioned previously in connection with contoured diagrams, the presence of a girdle developed in a vertical or near-vertical plane is an important guide to the parallel position, though at most stations it is not sufficiently well defined by 100 stones to be of diagnostic value. However, a rough correlation seems to exist between a strongly developed transverse orientation pattern and the presence of a girdle normal to the transverse position. These are logical consequences of deposition following rotational movement.

The criteria for distinguishing the parallel and transverse positions have been tested by applying them to actual cases. Results of the tests are given on the following pages. Figure 24 shows the regional relationships of the till-fabric data.

TESTS OF CRITERIA FOR DISTINGUISHING ORIENTATION POSITIONS

General statement.—A few stations in the central New York area yielded orientation diagrams that were anomalous or ambiguous. Hence these problem stations are ideal for testing the practical value of the criteria selected for determining the direction of glacier flow.

In the following analyses, no special weight is given to those indicators or guide forms that fall in more than one guide group. Each is recorded but once.

Station 15.—Station 15 is situated 0.6 mile northwest of Carpenter Pond, in the west central part of the Cazenovia quadrangle. The exposure is a road cut in one of the many till mounds on the west slope of a shallow through-valley. The rose pattern from this station is regular and symmetrical, and evidence from striae and topography is sufficient to show the relation of this pattern to the direction of ice

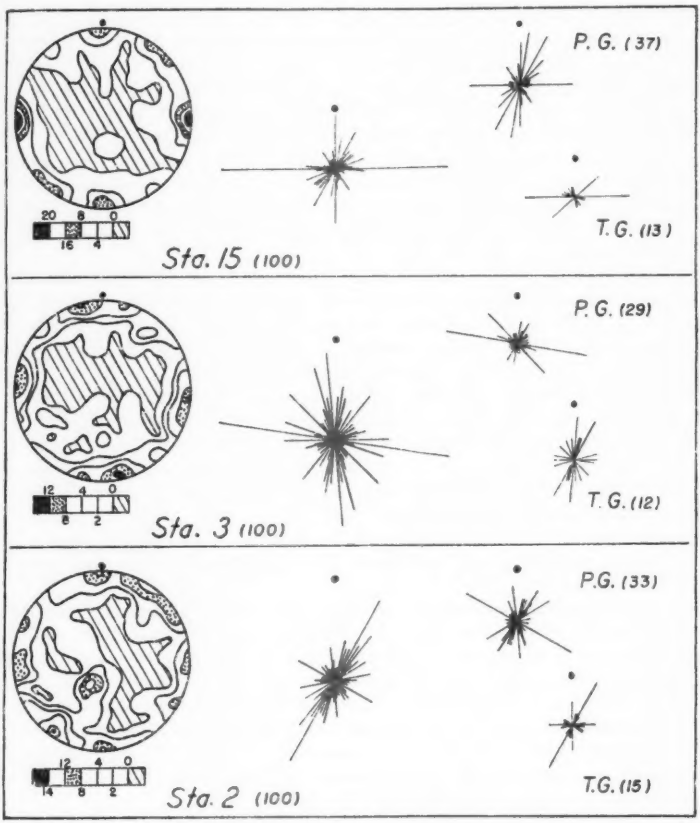

FIGURE 23.—*Analyses of anomalous patterns from three localities*

Parallel and transverse guide forms (indicated by P.G. and T.G., respectively) are shown separately. Their contribution to the complete station pattern indicates clearly the direction of glacier flow. True north is shown by a dot above each diagram.

flow, but the pattern is so different from most others that it was rejected in making up the composite group.

Figure 23 shows the various diagrams pertaining to this station. The contoured diagram shows an unusually strong east-west maximum, with almost no suggestion of a girdle across the diagram connecting the two parts of this maximum. However, an imperfect girdle is developed in a vertical plane normal to the east-west position. On the basis of these facts, the pattern is one of strong transverse orientation.

The guide forms are plotted separately. Each of the five stones oriented N. 10° E. is a guide to the parallel position, and two of them are included in more than one guide group. The average of intermediate-axis dips at this station is very low, and

the five parallel guides in transverse position appear there because of their low intermediate-axis dips. Even so, the statistical effect of these forms on the entire pattern is to show clearly that the stones oriented east-west are in transverse position. Therefore the analysis indicates that the orientation of stones at this station presents a strong transverse pattern. All field evidence confirms this interpretation.

Station 3.—Station 3 is located on the south bank of Limestone Creek, one mile northeast of New Woodstock (central part of Cazenovia quadrangle). The valley has a general east-west trend. Striae about half a mile to the southwest indicate ice movement due south; and about one mile northwest, striae are aligned S. 30° E. No evidence was found within the valley to show whether the ice flowed across it or along its axis.

The rose diagram of this station (Fig. 23) is characteristic except for its slight asymmetry. Though its longest line nearly parallels the valley trend, nothing indicates whether that line represents the parallel or the transverse position. The contoured diagram shows a slight tendency to general southward dip but offers no clear solution to the problem of ice-flow direction.

The parallel indicators show that the depositing ice flowed principally up-valley (eastward) but probably shifted its direction occasionally to a more southeasterly course. Seven of the eleven stones oriented N. 85° W. are good guides to the parallel position. Likewise, four of the six stones aligned N. 45° W. are parallel guides. The transverse indicators, though few in number, are in general agreement with evidence from the parallel guides.

The lithology of the till in this part of the valley changes abruptly about half a mile farther upstream (east), marking the limit of drift brought into the valley by ice from the northwest.

Station 2.—Station 2 is located on the south bank of Middle Branch Tioughnioga Creek, one mile southeast of Sheds (southeastern part of Cazenovia quadrangle). The valley slopes to the west, and the station is less than half a mile within the limits of drift brought from the northwest. This genetic relation of the till was not fully known at the time of field study, and the singularly symmetrical pattern (Fig. 23) was then interpreted as the work of ice flowing from the northeast, which had preceded the later flow from the northwest. But after the till relation was known, the diagonal position of the pattern with respect to the valley axis left considerable uncertainty as to the direction of glacier flow.

The contoured diagram suggests the presence of girdles other than the peripheral one, but not clearly enough for safe interpretation.

The indicators of the parallel and transverse positions show clearly that the pattern is one of prevailing transverse orientation. Five of the six stones aligned N. 60° W. are good parallel indicators, and the sixth is fairly good. Therefore the depositing ice was moving toward the southeast as it reached that part of the valley.

Conclusion.—Pattern analyses of these problem stations show that the guide forms give correct results in all cases where these results can be checked by other evidence. This gives assurance that the criteria are dependable where other evidence is lacking. Parallel-position guides are more numerous in an unselected group of 100 stones and give more decisive results than transverse-position guides. Decisiveness of results generally corresponds to the degree of simplicity and symmetry of the

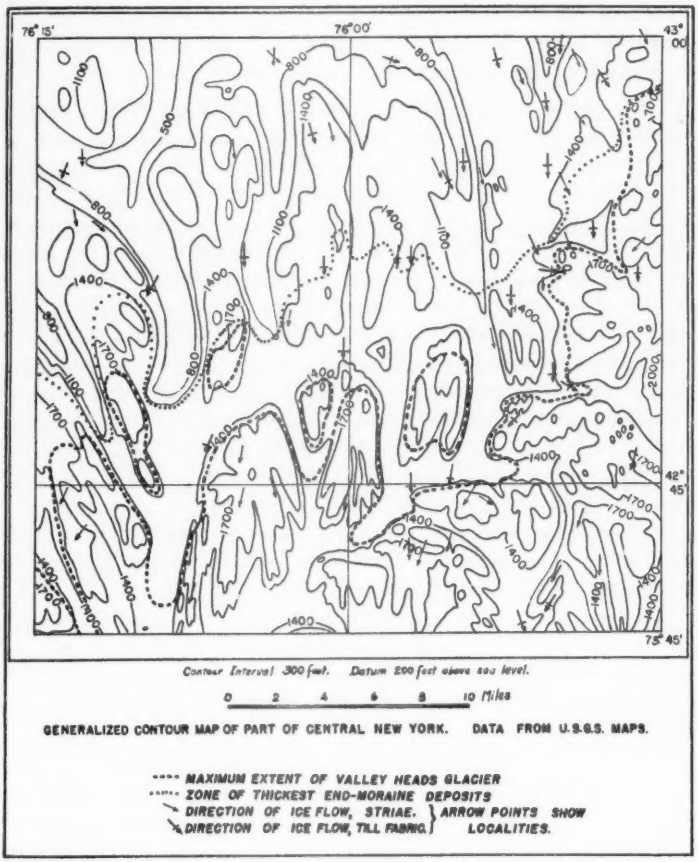
76° 15'
76°00'
43° 00'
42° 45'
75° 45'
Contour Interval 300 feet. Datum 200 feet above sea level.
0 2 4 6 8 10 Miles
GENERALIZED CONTOUR MAP OF PART OF CENTRAL NEW YORK. DATA FROM U.S.G.S. MAPS.
MAXIMUM EXTENT OF VALLEY HEADS GLACIER
ZONE OF THICKEST END-MORAINE DEPOSITS
DIRECTION OF ICE FLOW, STRIAE.
DIRECTION OF ICE FLOW, TILL FABRIC.
ARROW POINTS SHOW LOCALITIES.

FIGURE 24.—*Regional relationships of till-fabric data*

Cazenovia quadrangle (northeast), Tully quadrangle (northwest), and parts of Pitcher quadrangle (southeast) and Cortland quadrangle (southwest). The Valley Heads glacier moved southeast, and the earlier glacier toward the southwest. Strong local topographic control of ice-flow direction is apparent.

pattern to be analyzed. This correspondence seems primarily related to the constancy in direction of glacier flow during the time of deposition.

SUMMARY OF CONCLUSIONS

The conclusions drawn from this study of till fabric are believed to be valid for till in any region where the conditions of glaciation were com-

parable to those of central New York. Comparative studies from other regions seem highly desirable. The principal conclusions for the central New York area may be summarized as follows:

(1) The ground-moraine till accumulated gradually beneath the moving glacier, but with occasional local and temporary intervals of erosion that gave rise to boulder-pavement phenomena. Fabric patterns from successive layers of till record occasional shifts in direction of glacier flow similar to those shown by intersecting sets of striae on rock ledges.

(2) Glacier ice in which the stones were carried moved as a plastic solid flowing in obedience to the laws of fluid mechanics. Stones carried above the floor were subject to rotation about an axis lying in a plane of uniform shear and normal to the direction of glacier flow. Ordinarily the longest axis was the axis of rotation, though under some conditions, probably following temporary orientation by sliding, rotation about the intermediate or the shortest axis probably occurred. Stones rotating on or near the floor were subject to shoving into the deposited till by thrust from higher and faster moving debris. Many of the stones thus imbedded remained permanently in that position.

(3) Stones in contact with the glacier floor (or possibly along a well-defined shear plane in the ice) moved by sliding. Normally both long and intermediate axes were parallel to the floor plane, with the long axis aligned in the direction of movement. However, prominent flat surfaces, especially on wedge-form stones, commonly exercised more control in orientation by sliding than did the trend of the long axis. The controlling surfaces were aligned parallel to the direction of movement.

(4) Inherent characters such as form, roundness, size, and relative axial lengths, predispose a stone to certain kinds of movements and thus influence the likelihood of deposition in certain diagnostic attitudes or positions. Some types of stones have a relatively high statistical probability in depositional attitude, with the long axis aligned either parallel to or transverse to the direction of glacier flow. Thus it is possible to determine the direction in which the glacier moved while the deposit was accumulating. This method of ascertaining the direction of ice flow is independent of other evidence and can be used directly in the field.

(5) Minor differences in till-stone characters, such as slight rounding, have a profound effect on the depositional orientation preference of some types of stones (for example, Fig. 18). This in turn suggests that the number of such influences has not been fully measured and that a more detailed classification of till stones will probably reveal further signifi-

cant relations. Delicate adjustment to slight degrees of change is characteristic of glacial behavior, and the enormous forces of the continental glacier seem to have responded even to very small differences in those factors commonly grouped under the term structure.

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EXPLANATION OF PLATE 1

PLATE 1.—STONES FROM TILL IN CENTRAL NEW YORK

FIGURE 1. TYPICAL TILL STONES

Illustrating forms and degrees of roundness. About ⅛ natural size. All have remnants of original concave surfaces. Note low range of "*c*" roundness and nearly complete transformation of numbers 4 and 8 to ovoid form.

1	Tabular	"*c*"	10	Wedge-form	"*c*"
2		"*c*"	11		"*d*"
3		"*d*"	12	Discoid	"*b*"
4	Rhombohedroid	"*b*"	13	Ovoid	"*a*"
5		"*c*"	14		"*b*"
6		"*c*"	15		"*c*"
7		"*d*"	16	Varihedroid	"*c*"
8	Wedge-form	"*b*"	17		"*d*"
9		"*c*"			

FIGURE 2. WEDGE-FORM STONES

Showing striae parallel with prominent edges rather than with long axes of stones.

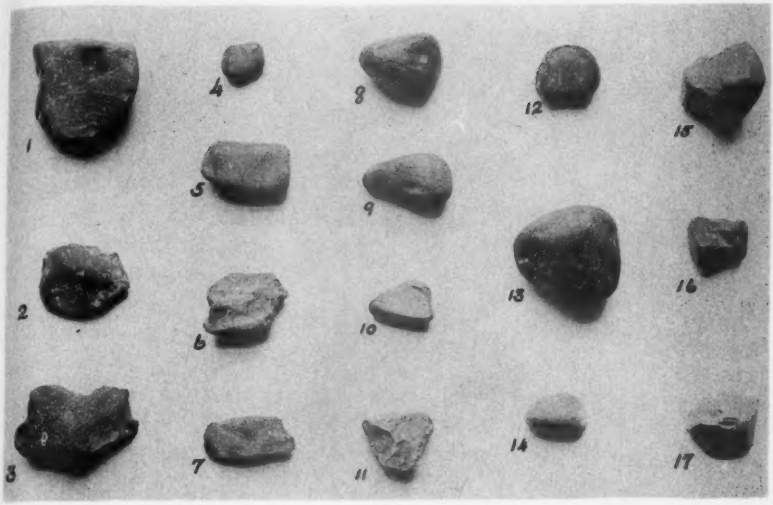


FIGURE 1

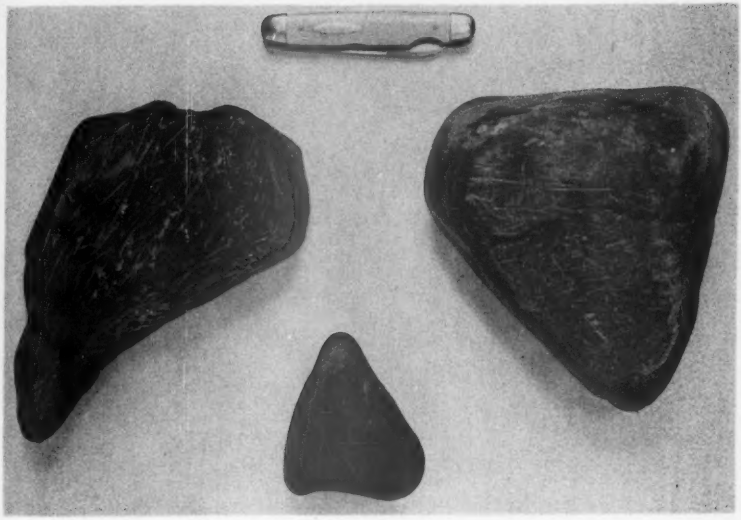


FIGURE 2

STONES FROM TILL IN CENTRAL NEW YORK

119°30'
R.27 E.
20'
R.28 E.
49°00'
48°50'
48°40'
UNITED STATES - CANADA
Kruger Mtn
Loomis
Driscoll I.
Williams I.
Lyhott I.
Mt Hull
LOOKOUT
Haley Mt
Wannacut Lake
Wauconda
Wauconda Mine
Crater Lakes
Whisky Mt.
Whitestone Mt.
Earl Mt
Antoine Valley
Lemanasky Mtn
Cayuse Mtn
Spring Coulee
Aeneas Lake
Tonasket
OKANOGAN
Bonaparte
Angelin
Mosquito Creek
Ninemile Creek
Siwash Creek
Antoine Creek
Bonaparte Creek
Aeneas Cr.
Omak Lake
Golden
Rainbow Mine
Okanogan Tree Boom Mine
QAL
TA
TS
MA
LA
SG
QD
OG
UA
WG
I
II
III
B.M.

LEGEND

INTRUSIVE ROCKS

MESOZOIC

SG — SIMILKAMEEN GRANODIORITE
(BIOTITE-HORNBLENDE GRANODIORITE, LOCALLY PORPHYRITIC)

WG — WHISKY MOUNTAIN GRANODIORITE
(PORPHYRITIC BIOTITE GRANODIORITE)

I II III — COLVILLE BATHOLITH

I GNEISSIC AND MYLONITIC FACIES.— BANDED MYLONITE & PROTOMYLONITE; GRANULITIC AND AMPHIBOLITIC GNEISSES; MIGMATITES; LARGE MASSES OF PEGMATITE NEAR MARGIN)

II SWIRLED AND PORPHYRITIC GNEISS FACIES.— (PROTOCLASTIC BIOTITE GNEISS, INTRICATELY SWIRLED; LOCALLY PORPHYRITIC WITH STRAIGHTER FOLIATION)

III HOMOPHANOUS FACIES.— (BIOTITE GRANODIORITE; HORNBLENDE-BIOTITE GRANODIORITE)

OG — OSOYOOS GRANODIORITE
(CRUSHED AND FOLIATED BIOTITE-HORNBLENDE GRANODIORITE)

QD — QUARTZ DIORITE
(FINE TO COARSE GRAINED EPIDOTIZED AND CHLORITIZED QUARTZ DIORITE AND DIORITE)

KM — KRUGER MOUNTAIN MALIGNITE
(HASTINGSITE-AUGITE-MICROCLINE MALIGNITE. LOCALLY SYENITIC)

MESOZOIC OR LATE PALEOZOIC

(DIKES)

METAGABBRO AND METADOLERITE
(CLINOZOISITE AMPHIBOLITE, "EPIDIORITE", CHLORITIZED DOLERITE AND GABBRO; PROBABLY IN PART CONTEMPORANEOUS WITH ANARCHIST VOLCANIC ROCKS)

BEDDED ROCKS

CENOZOIC

QUAT.

QAL — ALLUVIUM AND GLACIAL DEPOSITS
(MAINLY GLACIOFLUVIAL AND GLACIOLACUSTRINE SEDIMENTS)

UNCONFORMITY

EARLY TERTIARY

TA — ANDESITE FLOWS AND PYROCLASTICS
(PREVAILINGLY HORNBLENDE ANDESITE; SOME DACITE AND RHYOLITE. VOLCANIC NECKS NOT DIFFERENTIATED)

TS — SEDIMENTARY ROCKS
(GRANULE TO BOULDER CONGLOMERATE, ARKOSE, SHALE; MAINLY OF FLUVIATILE ORIGIN)

UNCONFORMITY

TERTIARY?

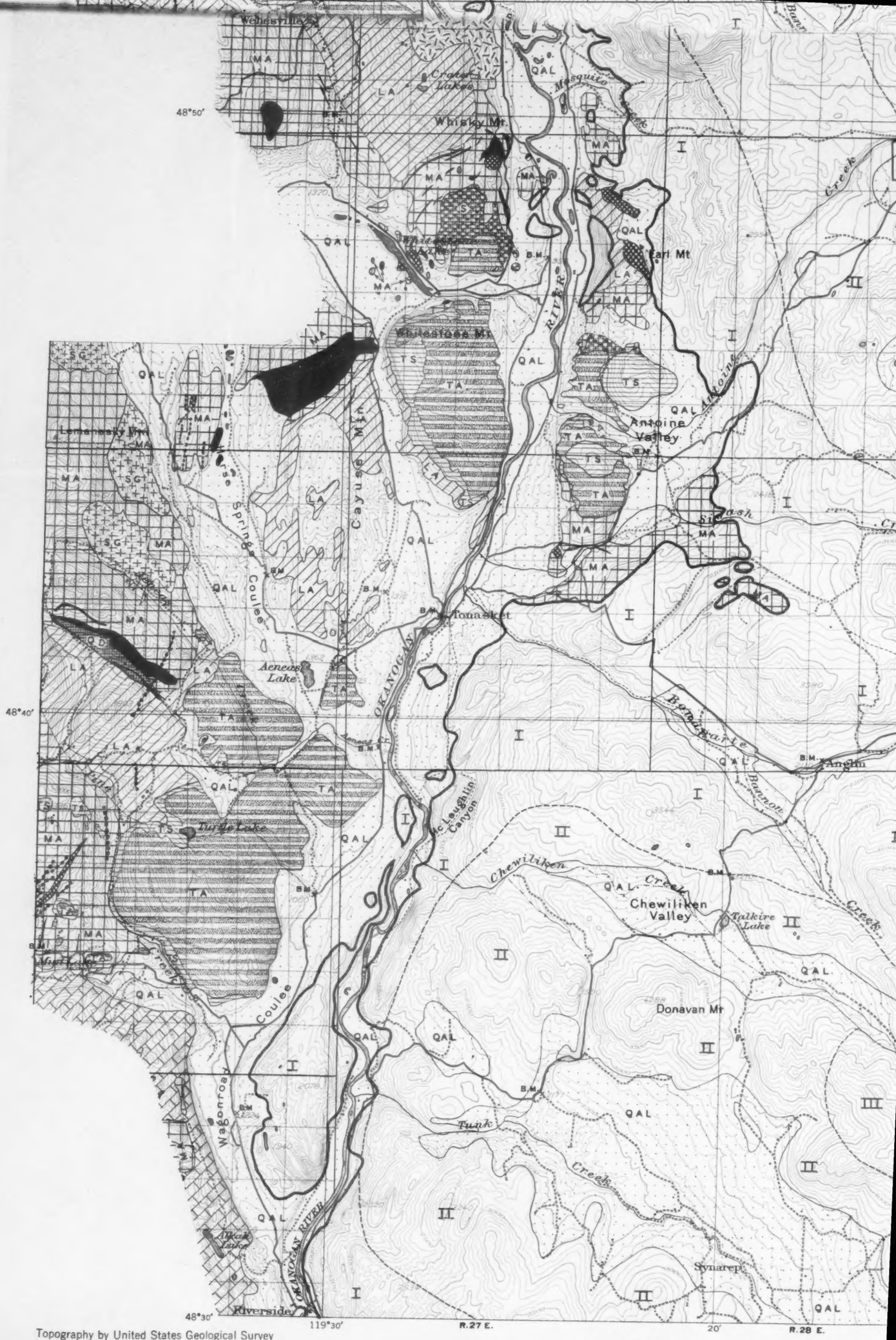
S — SPILITE
(SPILITE AND KERATOPHYRE)

UNCONFORMITY

TRIASSIC

LS — LIMESTONE
(MASSIVE LIMESTONE, LOCALLY SILICEOUS, LOCALLY DOLOMITIC. THIN DISCONTINUOUS CONGLOMERATE BED AT BASE)

QD



Topography by United States Geological Survey

GEOLOGIC MAP OF PART OF BORDER O

DOLOMITIC. THIN DISCONTINUOUS CONGLOMERATE BED AT BASE)

QD

QUARTZ DIORITE
(FINE TO COARSE GRAINED EPIDOTIZED AND CHLORITIZED QUARTZ DIORITE AND DIORITE)

KM

KRUGER MOUNTAIN MALIGNITE
(HASTINGSITE-AUGITE-MICROCLINE MALIGNITE. LOCALLY SYENITIC)

MESOZOIC OR LATE PALEOZOIC

(DIKES)

METAGABBRO AND METADOLERITE
(CLINOZOISITE AMPHIBOLITE, "EPIDIORITE", CHLORITIZED DOLERITE AND GABBRO; PROBABLY IN PART CONTEMPORANEOUS WITH ANARCHIST VOLCANIC ROCKS)

BEDDED ROCKS

CENOZOIC

QUAT.

QAL

ALLUVIUM AND GLACIAL DEPOSITS
(MAINLY GLACIOFLUVIAL AND GLACIOLACUSTRINE SEDIMENTS)

UNCONFORMITY

TERTIARY

EARLY TERTIARY

TA

ANDESITE FLOWS AND PYROCLASTICS
(PREVAILINGLY HORNBLENDE ANDESITE; SOME DACITE AND RHYOLITE. VOLCANIC NECKS NOT DIFFERENTIATED)

TS

SEDIMENTARY ROCKS
(GRANULE TO BOULDER CONGLOMERATE, ARKOSE, SHALE; MAINLY OF FLUVIATILE ORIGIN)

UNCONFORMITY

TERTIARY?

S

SPILITE
(SPILITE AND KERATOPHYRE)

UNCONFORMITY

TRIASSIC

LS

LIMESTONE
(MASSIVE LIMESTONE, LOCALLY SILICEOUS, LOCALLY DOLOMITIC. THIN DISCONTINUOUS CONGLOMERATE BED AT BASE)

UNCONFORMITY

LATE PALEOZOIC

UA

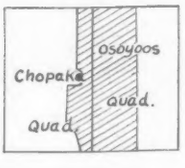
UPPER ANARCHIST SERIES
(PREDOMINANTLY METAANDESITE AND METABASALT, IN PART SPILITIC. COMMONLY MASSIVE GREENSTONE, IN PART WELL FOLIATED CHLORITE SCHIST, SOME GRAY PHYLLITE AND FINE QUARTZITE.)

MA

MIDDLE ANARCHIST SERIES
(METACONGLOMERATE, QUARTZITE, GREYWACKE, LIMESTONE, METAVOLCANIC ROCKS (ESPECIALLY IN NORTH), LOCAL PHYLLITE. LIMESTONE PODS ARE SHOWN IN SOLID GREEN.)

LA

LOWER ANARCHIST SERIES
(PREDOMINANTLY PHYLLITE BUT CHLORITE SCHIST, FINE GRAINED QUARTZITE, AND SCHISTOSE GRANULE CONGLOMERATE LOCALLY ABUNDANT. ALL ROCKS COMPLEXLY SHEARED AND CRUMPLED.)



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ER OF COLVILLE BATHOLITH

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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PROTOCLASTIC BORDER OF THE COLVILLE BATHOLITH

BY A. C. WATERS AND KONRAD KRAUSKOPF

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ABSTRACT

In the Okanogan Highlands of Washington a complex plutonic mass called the Colville batholith intrudes folded and dynamometamorphosed sedimentary and volcanic rocks of late Paleozoic and Triassic age. Along the sharply discordant contact the wall rocks are much fractured and granulated, but contact metamorphism is slight or absent.

The batholith is remarkably heterogeneous both structurally and petrographically. A central mass of structureless granodiorite grades outward into a belt of foliated igneous rock which commonly shows intricate swirling of the foliation. These swirled rocks grade into a peripheral belt of variable but well-foliated migmatitic gneisses characterized by severe granulation of the constituent minerals. Over broad zones this rock is a mylonite; locally recrystallization has produced types resembling metamorphic granulites. That the crushing was protoclastic and not due to dynamic metamorphism following batholithic solidification, is proved by the relations with the wall rocks and by the widespread cementation of the broken materials by films and stringers of undeformed quartz and microcline.

Along the contact between the approximately contemporaneous Osoyoos and Colville batholiths occurs a narrow belt of heterogeneous syenite with highly complicated internal structure. This is believed to be a hybrid rock formed by the action of magmas and emanations from both batholiths upon a thin wall rock septum.

Deep erosion revealed the Colville rocks over wide areas. Flows of spilite and keratophyre were poured out, and after another erosion interval were covered by Tertiary sediments and volcanics. These were then folded and deeply eroded. Ice sheets covered the area during the Pleistocene.

INTRODUCTION

In the Okanogan Highlands, the region between the Okanogan River and the Columbia River is underlain mainly by a large plutonic mass, named the "Colville granite batholith" by Pardee (1918, p. 30). In view of the remarkable heterogeneity and complex history of this intrusive the term batholith may seem objectionable, though the very general definition of batholith advocated by the Batholith Committee of the National Research Council (Grout, *et al.*, 1933, p. 2) would apply. However, petrographic study shows that granite is very subordinate; granodiorite, quartz diorite, and other rock varieties are much more abundant. Because of its petrographic and structural heterogeneity no single rock term is appropriate to it. Therefore, the intrusive will be called the "Colville batholith" in this report.

The structural heterogeneity of the Colville batholith was appreciated by Pardee, who described the border of the mass as follows (p. 33):

> "The most noticeable and extensive local structural feature in the Colville granite is a schistose or gneissic foliation . . . so plainly developed that outcrops seen at a distance suggest a thin-bedded sedimentary rock . . . The feldspar phenocrysts are slightly distorted and the groundmass shows shearing strains."

The remarkable character of this border zone attracted the attention of Waters during a reconnaissance in 1931. Opportunity for detailed examination was afforded during the summers of 1935 and 1936 when the senior author mapped the southwest quarter of the Osoyoos quadrangle and adjoining parts of the Chopaka quadrangle. Krauskopf extended the mapping to the northwestern quarter of the Osoyoos quadrangle in 1936 and 1937.

ACKNOWLEDGMENTS

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GEOLOGIC SETTING OF THE COLVILLE BATHOLITH

DISTRIBUTION OF PLUTONIC ROCKS

The Colville batholith is one of several closely related intrusions in northern Washington, Idaho, and adjacent parts of British Columbia. Many of them have not been mapped or even investigated in reconnaissance fashion throughout their entire extent. Their age of emplacement is not definitely known, though most of them cut Paleozoic and

early Mesozoic strata and are overlain by, or have contributed detritus to, sediments of Tertiary age. They are generally assumed to have been emplaced in the late Jurassic or early Cretaceous, and to be associated with the widespread igneous activity that accompanied the Nevadan

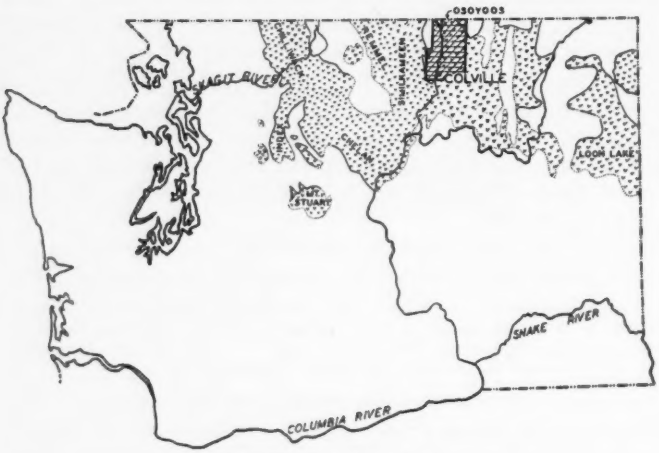


FIGURE 1.—*Index Map*

Mesozoic intrusive rocks shown by carets. Solid lines, intrusive contacts; dotted lines, unconformable contacts. Names applied to various intrusive masses are indicated. Area covered by Plate 1 shown by cross-hatching.

Revolution. The position of the Colville batholith with relation to other large plutonic masses is indicated on Figure 1, which has been modified from maps published by the Department of Conservation and Development, State of Washington, and by the United States Geological Survey.

Figure 1 also indicates the part of the Colville batholith studied in detail for this report. The area mapped constitutes only a small fraction of the total surface area of the intrusion, and it is possible that relations observed in this small marginal area may not hold for the entire mass. Reconnaissance examination of much of the batholith indicates, however, that the features observed in the mapped area in Okanogan Valley are duplicated at many other points along the periphery. Campbell (1940) has recently found similar relations along the east border of the mass in the Kettle Range.

ROCKS OLDER THAN THE COLVILLE BATHOLITH

The Anarchist series.—In the Okanogan Valley region the Colville batholith is intrusive into a varied group of rocks whose structural and

stratigraphic relations are only imperfectly known. The oldest and most widespread unit is a heterogeneous group of greenstones, phyllites, quartzites, chlorite schists, crystalline limestones, and partially metamorphosed conglomerates which Daly (1912) named the Anarchist series. This series can be subdivided into many lithologically distinct formations, although mapping is difficult because several lithologic varieties appear at more than one horizon. Long ago crumpled and metamorphosed, later invaded by one great batholith after another, still later subjected to deep erosion, these ancient rocks are today so contorted and so fragmentary in distribution that their origin and initial structures are almost undecipherable. Foliation is nearly universal, only a few resistant quartzites and greenstones having escaped. Often all trace of the original bedding has disappeared; where visible it usually parallels the schistosity. Sedimentary structures, which might give some hint of the top and bottom of beds, have been almost entirely obliterated.

Because of the complex structural relations of these rocks and the absence of fossils, it is not always possible to work out completely the areal distribution of each formational unit. Therefore, on the geologic map (Pl. 1) only a very broad threefold subdivision of the Anarchist series has been attempted, and instead of introducing new names for these units they have been called the Upper Anarchist, Middle Anarchist, and Lower Anarchist. A similar subdivision of the Anarchist rocks had earlier been attempted by Umpleby (1911, p. 65).

The total thickness of the Anarchist series is not known but is undoubtedly very great. Sections over 5000 feet thick, apparently without duplication, include only a part of one of the three divisions of the Anarchist.

Lower Anarchist: The Lower Anarchist consists predominantly of black, lustrous phyllites. Near the batholithic masses recrystallization has been more profound, and mica schists appear. Locally metamorphism has not advanced beyond the stage of slate. The phyllites are mainly quartz-sericite rocks with feldspar scarce or absent. However, the phyllites locally grade into types probably derived from tuffaceous material, and in these the content of feldspar (albite), chlorite, and epidote is relatively high. These rocks are commonly associated with true chlorite schists which represent regionally metamorphosed mafic lavas and tuffs. In general shearing and recrystallization has all but obliterated original textures, though locally resistant nodules preserve relict volcanic textures.

Although phyllites and chlorite schists are most common in the Lower Anarchist, other rock types are found. The more arenaceous varieties of phyllite grade locally into true quartzites. South and west of Whitestone

Mountain the Lower Anarchist is composed largely of schistose conglomerate. Although conglomeratic structure is clearly preserved, the original pebbles (which consisted mainly of quartz and siliceous rock fragments) have been thoroughly sheared, granulated, and stretched, and the matrix has recrystallized into coarse foils of mica and quartz which give the rock a pronounced foliation. Most of these metamorphosed conglomeratic rocks were originally granule conglomerates, though in a few beds the pebbles attain a diameter of 2 to 4 inches.

The rocks here designated as Lower Anarchist are well exposed on the central and southern slopes of Whisky Mountain (Pl. 1), in the highly crumpled axial portion of a broad anticline defined clearly by the more competent Middle Anarchist strata. The anticline plunges to the southeast, and its northern limb is invaded by an elongate stock of granodiorite. Southwest of this area Lower Anarchist rocks are widely exposed in the Cayuse Mountains and in the plain separating the Cayuse Mountains from Whitestone Mountain. In this region the chlorite schist and schistose conglomerate facies are exceptionally well developed. Still farther southwest this belt of Lower Anarchist rocks is reduced (Pl. 1) to a narrow band in the mountains southwest of Horse Springs Coulee. It then expands into a broad area of chlorite schist and phyllite in the mountains at the headwaters of Pine Creek. Despite the thorough crumpling of the incompetent Lower Anarchist rocks certain bands of chlorite schist can be traced continuously enough to suggest that in the vicinity of upper Pine Creek these rocks probably form an anticline whose southwestern limb is overturned. This interpretation is strengthened by the synclinal structure of the belt of Middle Anarchist rocks which separates the area of Lower Anarchist in Pine Creek from that in Horse Springs Coulee and the Cayuse Mountains.

South of the Pine Creek area Lower Anarchist rocks have not been recognized with certainty, although they may be represented in the highly complex region just southwest of the mapped area.

Middle Anarchist: The Middle Anarchist is more heterogeneous than the Lower Anarchist. Although phyllites are locally common, the metamorphic derivatives of sandstones and conglomerates are more widespread. Metamorphic limestones are also common but peculiar in distribution. They usually occur not as distinct widespread layers but as unusual pods and lenses which may be of considerable thickness but are very limited in horizontal extent. Inasmuch as the lensing is quite unrelated to faulting or stretching during deformation, the podlike structure of the limestone must be an original feature, formed perhaps as the result of deposition in reefs. Metamorphism has generally obliterated all fossil remains, but in a few places crinoid and coral fragments were

found. The limestone pods of large size have been specially designated on Plate 1. Near the lower part of the Middle Anarchist section the limestones are commonly associated with quartzitic and phyllitic rocks. Higher in the section they occur with quartzites, metamorphosed conglomerates, and metamorphosed lavas.

Metamorphic derivatives of conglomerates and sandstones are the most common varieties in the Middle Anarchist. North and west of Horse Springs Coulee the basal part of the Middle Anarchist consists largely of quartzite with occasional interstratified bands of black phyllite. In the lower part of the section the quartzites are gray to black and are composed almost entirely of quartz with a little graphite and muscovite. Locally the beds are bluish and so fine-grained as to resemble chert. Higher in the section these pure quartzose rocks give way to more feldspathic varieties associated with siltstones and thin lenses of conglomerate. Just north of Mud Lake beds of this character are intercalated with thin-bedded black limestones which are rarely more than 2 or 3 feet thick, but which have great lateral extent. In their broad horizontal extension and their black color these limestones are sharply differentiated from the typical podlike masses of white to blue-gray limestone forming the more common calcareous representatives of the Middle Anarchist.

On Lemanasky Mountain and east of Lemanasky Lake strata believed to be still higher in the Middle Anarchist series crop out. Here in addition to the prevailing quartzites much of the section is composed of boulder and pebble conglomerates interstratified with medium-grained sediments which, due to the abundance of dark-colored rock fragments, are best described as graywackes. The fragments composing both graywackes and conglomerates are chiefly mafic volcanic rocks, though locally beds made largely of fragments of black chert, vein quartz, and other materials are present.

Although the pebbles of these conglomerates have been sheared to some extent and are rather thoroughly recrystallized, they do not show the profound mashing and foliation characterizing the conglomerate beds of the Lower Anarchist. This does not necessarily mean that the Lower Anarchist was metamorphosed before the Middle Anarchist was laid down but is more probably explained by the marked difference in competency between Middle Anarchist and Lower Anarchist rocks. It is significant that the best preserved fossils in the Anarchist series have been recovered from the tough graywackes of the Middle Anarchist.

Although volcanic fragments are common as detrital material in Middle Anarchist rocks from the southern part of the area, true metamorphosed lavas are absent or very infrequent. To the north, however, chlorite schists are abundant, becoming the dominant rock of the Middle

Anarchist just beyond the northeast corner of the mapped area. Limestone pods are still numerous, and the characteristic massive quartzites, graywackes, and metamorphosed conglomerates persist in subordinate amount. The transition between a dominantly sedimentary sequence and one with abundant volcanic material is well shown around the nose of the anticline on the southeast slopes of Whisky Mountain. Southwest of the anticlinal nose appear a few light-colored, quartzose chlorite schists (probably derived from tuff beds) interbedded with phyllite, limestone, and quartzite. Around the nose to the north the proportion of chlorite schist becomes steadily larger, the beds growing more numerous and more massive. Because the transition takes place gradually along the strike, it seems best interpreted as a change of facies in the Middle Anarchist.

The relations between the Lower Anarchist and Middle Anarchist are believed to be conformable, although in such structurally complex metamorphosed rocks the presence of disconformities, or even of structural unconformities with considerable discordance might easily go undetected. Conformability is suggested at several localities by the dark phyllites of the Lower Anarchist becoming sandier and not clearly separable from the black quartzites at the base of the Middle Anarchist. Drawing a contact between the two series in such localities is largely done arbitrarily. Locally relations seem to suggest a strong angular discordance, but on closer study are found to be the result of strong deformation. One such locality is near the northern end of Cayuse Mountain. A large pod of white limestone near the base of the Middle Anarchist crosses the mountain near its north end. Though locally contorted and variable, the layers forming this pod have an average strike of about N. 50° E. and dip northwest at angles of 20 degrees to 50 degrees. A short distance south of the limestone, chlorite schists and phyllites of the Lower Anarchist are extensively exposed. They are highly crumpled, but in the central part of Cayuse Mountain their prevailing strike is about north-south with steep dips to the east; the strike even turns to the northwest in the southern and western parts of the mountain. These rocks therefore appear to strike into and beneath the limestone pod almost at right angles. However, in a narrow zone immediately beneath the limestone pod the underlying phyllites are highly contorted. At the contact with the limestone they have usually swung around to approximate parallelism with the overlying bed. It appears that during deformation the highly incompetent phyllites and chlorite schists were minutely folded along planes roughly at right angles to the regional strike. However, the overlying strong limestone did not yield to these minor contortions, and has sheared more or less bodily over the weak rocks. Significantly, the axes of plications in these incompetent materials practically always maintain the

same pitch as the dip of the massive limestone above. Such relations are common in regions of mildly metamorphic rocks of varying competency; excellent examples have been described and figured by Balk (1932). In the Anarchist rocks such relations are seen not only at the junction between the Lower Anarchist and Middle Anarchist, but also at the base of thick limestone pods and similar layers within the Middle Anarchist. Southeast of Lemanasky Lake the zone of weakness formed beneath a limestone pod as a result of this kind of deformation has served to localize the position of a large quartz diorite sill.

Upper Anarchist: The Upper Anarchist is composed predominantly of metamorphosed volcanic rocks, with some interbedded phyllites and fine quartzites. The best exposures of Upper Anarchist rocks are in the northeast corner of this area, where a great variety of green chloritic rocks abound. Some are soft, finely foliated, often minutely crumpled chlorite schists. At the other extreme are little deformed lavas, with outlines of pyroxene and amphibole phenocrysts still distinctly visible, though the microscope reveals complete breakdown of the parent minerals. Occasional coarser-grained altered mafic dikes are associated with the extrusive material.

The Upper Anarchist greenstones are interesting chiefly for their high albite content. This is the only feldspar observed, and it commonly constitutes over half the volume of the rock. Even in the least-altered lavas and coarse-grained dikes, where seemingly original sharp-edged, feldspar shapes are preserved, the feldspar is entirely albite (An_{05}-An_{10}). The abundance of sodic plagioclase suggests derivation from spilitic lavas, a suggestion borne out by the 6.04 per cent of soda in the single analysis available.

Upper Anarchist rocks are believed to lie conformably over the Middle Anarchist series, although difficulties in interpreting the relations arise because of crumpling near the contact. The best contact exposures are just west of the limits of this area on the eastern side of Palmer Mountain, and in the northeast corner of the area just northeast of the headwaters of Tonasket Creek. In these two localities relations seem to be strictly conformable, and boundaries between the two series can be drawn only arbitrarily. For a short distance just east of Kruger Mountain lavas of the Upper Anarchist apparently overlie the upturned Middle Anarchist beds, but this may well be due to the massive nature of the lavas and the local relative incompetency of the Middle Anarchist.

Age of Anarchist series: Structures of organic origin were found at many places in the Anarchist series, but were almost invariably so metamorphosed as to preclude specific identification. However, in the resistant graywackes of the southwestern part of sec. 4, T. 38 N., R. 27 E.,

at a point about 100 yards south of the small saline which is being worked for epsomite, fossils were found which, though poorly preserved, showed many of the original morphological details. These fossils were kindly examined by the late Dr. George Girty who reported on them as follows (personal communication):

"The collections appear to represent the same fauna and horizon. The fossils in these collections are flattened and distorted and many of them are fragmentary. It is impossible to advance even generic identifications with confidence. The fauna is primarily noteworthy for its abundance of gastropod shells, but the rock almost invariably breaks across these fossils instead of around them. Some very long and slender shells are apparently shaped like *Loxonema* or *Zygopleura*, others like *Orthonema*, and many more or less like *Natica* or *Naticopsis*, that is, they appear to have a low spire and consist only of a few rapidly enlarging volutions. In fact, owing to distortion and the direction in which fractures pass through them, it is rarely possible to tell whether they represent a spirally coiled form, like *Natica*, or one coiled in one plane, like *Bellerophon*. Several good sized specimens are unquestionably coiled in one plane and at the same time are not chambered. I believe them to be Bellerophontids, probably a large specimen of *Euphemites*. To represent the pelecypods there is one good specimen which seems to belong to the genus *Myalina* and to be related to *M. subquadrata*, but *Myalina* occurs in the lower Triassic as well as the Carboniferous. The brachiopods are extremely rare, but there are several specimens that suggest a large species of *Chonetes*, unstriated and completely flattened. There are also several rugose corals (or what appear to be such) and several round crinoid stems . . . . I note also a small Spiriferoid (*Spiriferina*?) and ornamented shell fragments that might belong to *Productus* and also might belong to many other types. One impression suggests a bryozoan (*Fenestella* or *Polypora*) but might be part of some shell with cancellated sculpture.

Taking the evidence all in all, and realizing that much of it is unreliable, I believe that the age of these collections is probably Carboniferous and most likely Permian."

Subsequently another lot of fossils was obtained from this locality, but were in an even worse state of preservation. After examining them Dr. J. B. Reeside wrote (personal communication) as follows:

"The conspicuous things in it are a high-spired gastropod that may be what Dr. Girty compared to *Loxonema* and *Zygopleura* and a low-spired, stout gastropod that may be what he compared to *Natica* and *Naticopsis*. I did not note any bellerophontids, brachiopods, or bryozoans. The few pelecypods are too poor to mention. Taken by itself I would not want to assign an age to it, but I suspect very strongly that it belongs with . . . [the lot Dr. Girty examined] and that it is Permian, if they are."

To the north of this area in southern British Columbia occur rocks which have yielded definite Permian fossils and which are lithologically similar to the Anarchist series.

Limestone west of Wagonroad Coulee.—The region south of Mud Lake Valley and west of Wagonroad Coulee is underlain principally by massive limestones and dolomites hitherto undescribed in the literature. Although they form an extensive stratigraphic unit, they have so little bearing on problems connected with the Colville batholith that a study in sufficient detail to justify their naming and description as a separate formation was deemed unnecessary for this report. Such a study

would be both difficult and time-consuming, as structures in this region are highly complex.

Tentative conclusions as to the history of this area are as follows: After deposition, the Anarchist series was complexly folded and probably partially metamorphosed. Following a long erosion period resulting in peneplanation, the area again went beneath sea level and the limestones seen in the bold cliffs on the west side of Wagonroad Coulee were laid down. The sea seems to have been very clear, for aside from a few feet of basal conglomerate which intervene directly at the contact in most places, pure limestones and dolomitic limestones were deposited directly upon the truncated folds in the Anarchist series. The basal conglomerate establishes beyond any doubt that these limestones are actually unconformable above the Anarchist. The limestones are not pods like those in the Middle Anarchist, beneath which the more incompetent rocks have been crumpled and sheared until a semblance of an unconformable relation resulted, for the basal conglomerate is a true conglomerate and not a fault breccia. It is composed of fragments (generally well rounded) from the underlying rock types, and never of fragments of the limestone above. Although as a rule fragments of the variety of rock immediately underlying the limestone are most common, the conglomerate does contain pebbles of other Anarchist rock varieties which must have been carried some distance, as well as pebbles not indigenous to this area. Furthermore, good stratification can locally be seen in the conglomerate, thus distinguishing it unequivocally from fragmental material formed by fault brecciation. In places, the conglomerate is somewhat sheared and brecciated—evidently resulting from differential movement between the strong competent limestones above and the weaker Anarchist rocks below, but at several points the conglomerate is undeformed. Excellent exposures of this basal conglomerate as well as the surface of unconformity can be studied in the south central part of section 4, and near the center of sec. 9, T. 35 N., R. 26 E. These relations are also clearly seen along the eastern margin of sec. 32, T. 36 N., R. 26 E.

The limestone above the basal conglomerate is commonly massive and very thick. It may be well or indistinctly bedded. In the lower 500 feet of the section stratification is commonly indistinct or absent. Much of the rock is dolomitic, light-colored, fine-grained, and slightly siliceous, but in other parts of the area darker-colored, coarsely crystalline, well-bedded limestones prevail, and the exact sequence of these different lithologic varieties has not been carefully worked out. Calcareous siltstones and calcareous quartzites are rare and are usually thin.

Just east and south of Alkali Lake the prevailing massive white limestone is overlain by a gray to buff limestone containing considerable sandy material and passing locally into calcareous sandstone. In these rocks identifiable fossils were found.

The best preserved fossils come from a sandy limestone member extensively exposed in several hills and slopes near the southern end of Alkali Lake. One of the best collecting localities is in the northwestern part of sec. 26, T, 35 N., R. 26 E. Fossils from here were examined by Dr. J. B. Reeside who reported on them (personal communication) as follows:

"The lot . . . is a purely pelecypod fauna and of a type I have seen only in the Lower Triassic. There are probably several species of *Aviculopecten* represented by many individuals, a *Pinna*-like form, a monotid, a form that could be called *Pleuromya,* and a lot of small individuals that lack details. One could not without reservation say that this is a Mesozoic fauna, because similar types also occur in the late Paleozoic. The complete absence, however, of anything definitely Paleozoic and the general resemblance to definitely Triassic faunas elsewhere make me think that it is Mesozoic and probably Triassic."

Near Amy Lake, which lies about 1 mile to the northwest, occur beds similar to those at the above locality. Here Ralph J. Roberts and Warren Hobbs found the remains of a large ammonite which Dr. Reeside considered to be Triassic or Lower Jurassic.

Intrusive rocks older than Colville batholith.—Much of the structural complexity of the Anarchist series is due to the invasion of these rocks by intrusives of various kinds. In fact the Anarchist area of the Okanogan Valley represents merely a small wedge-shaped basin of metamorphic rocks preserved as a septum between three great batholithic masses—the Colville batholith which makes up the Colville and Kettle ranges on the east, the Similkameen batholith which underlies the Okanogan Range to the west and southwest, and the Osoyoos batholith, which begins a few miles south of the International Boundary and extends northward into the Okanogan Valley of Canada.

In addition to the batholiths, the Anarchist series has been riddled by many different varieties of plutonic and hypabyssal rocks which crop out in smaller masses. Several different periods of igneous invasion are involved. Some of the plutonic masses appear to have formed during the time when the Anarchist series was being deposited and may well represent the feeders through which the abundant volcanic rocks in the Middle and Upper Anarchist series rose to the surface. Others were emplaced later, but sufficiently early to be deformed and metamorphosed with the Anarchist rocks. Still others are definitely younger than much of the deformation and metamorphism of the Anarchist, though they have been cut by satellitic dikes from the great batho-

lithic masses and altered by contact metamorphism at batholithic contacts. Youngest of all are andesite porphyries probably representing feeders of the Tertiary volcanic rocks that crop out at various points along the Okanogan Valley. The exact number of igneous sequences has not been determined, but at least four and probably more periods of invasion are recorded.

METAGABBRO AND METADOLERITE: Among the oldest of these metamorphic rocks are masses which were once intrusions of gabbro and dolerite, but were subsequently thoroughly recrystallized. These altered gabbros are well exposed on Earl Mountain, in the low hills east of Tonasket, and at the south end of Whisky Mountain. Much larger, though somewhat less profoundly metamorphic bodies of gabbro occur on Palmer Mountain and Elemeham Mountain just west of the mapped area. Metamorphosed dolerites occur as sills and dikes at many localities; the most conspicuous ones are indicated in Plate 1.

The metagabbro in the low hills east of Tonasket has bright lustrous, curiously mottled feldspars. The feldspars are intergrown with aggregates of black amphibole. Microscopically the amphibole proves to be uralitic, clearly pseudomorphic after pyroxene. The feldspar is albite, and the mottling is due to numerous enclosed blebs of clinozoisite. This rock was originally a gabbro, but its labradorite has broken down into albite and clinozoisite and its pyroxene has been transformed to amphibole. The original gabbroic texture is excellently preserved.

The mass on Earl Mountain is similar save that it has suffered a wave of potash metasomatism from a large pegmatitic mass related to the Colville batholith. Emanations from this pegmatite have largely replaced the feldspars with microcline and have induced the formation of biotite.

Many metagabbros are unlike those just described in that their original textures have been completely obliterated by shearing and mashing prior to and during recrystallization. These sheared and recrystallized gabbros are represented by many different varieties of metamorphic rocks. Clinozoisite amphibolites, amphibolite schists, and epidotic granulites are perhaps most common.

The metamorphosed equivalents of the dolerites are similar. Many are mildly schistose greenish rocks still preserving traces of the original igneous texture though composed mainly of amphibole, sodic plagioclase, and epidote or clinozoisite. They are similar to the rocks called "epidiorites" in Scotland. Other dikes and sills grouped as metadolerite apparently belong to a younger period of igneous invasion and have not been so thoroughly metamorphosed. They show moderate shearing with the development of chlorite and the clouding and sericitization of feld-

spars, but without the profound recrystallization characteristic of the gabbros and dolerites previously described.

Some of the masses grouped as metagabbro and metadolerite in Plate 1 are really metamorphic equivalents of pyroxenites and other ultramafic rocks. In general they show abundant development of amphibole with or without garnet.

QUARTZ DIORITE: Near the headwaters of Pine Creek a somewhat irregular sill-like body has been injected into a plane of weakness caused by the shearing of a large pod of Middle Anarchist limestone over crumpled and contorted phyllites and chlorite schists. Although this mass is somewhat variable in composition it is dominantly a quartz diorite rich in hornblende. Apparently it was subjected to some of the movements which deformed the adjoining rocks and because of this partial metamorphism is believed to be older than the near-by batholithic masses otherwise resembling it in composition.

Petrographically the rock is composed of plagioclase (An_{35}) hornblende, quartz, and epidote named in order of decreasing abundance. Sphene, magnetite, and apatite are accessories. A very little pyroxene (diopside?) is found at the core of some hornblende crystals. Most of the minerals have been somewhat shattered by mechanical movements and chlorite, epidote, quartz, and untwinned albite have grown abundantly in the shear zones. In certain areas epidote makes up as much as one third of the rock. Locally the rock is stained by films and crusts of malachite.

This small mass seems to have accomplished a surprising amount of metamorphism. The rock for several tens of feet from the contact is locally impregnated with considerable quartz and epidote, and parts of the adjacent limestone have been transformed to lime silicate rocks.

The quartz diorite closely resembles an igneous body lying directly west of Osoyoos Lake, which Daly (1912, p. 439) considered to be a part of the Osoyoos batholith. Whether this mass west of Osoyoos Lake is really to be correlated with the large Osoyoos batholith east of the lake is a problem on which satisfactory information could not be obtained.

KRUGER MOUNTAIN MALIGNITE: Exposures of this rock are limited to two small hills on either side of the Okanogan Valley about 4 miles south of the International Boundary. The rock is considered to represent a part of the Kruger Mountain malignites which are best developed just northwest of this area. It is a dark, coarse-grained rock, consisting dominantly of the three minerals augite, hastingsite, and microcline. A little andesine is usually present, and albite occurs in microperthitic intergrowths, as rims around other feldspars, and in irregular patches. No

nepheline was observed, although its decomposition products may be represented by patches of fine micaceous material.

The malignite has no gneissic structures, but on the hill east of Okanogan Valley it shows the effects of extreme cataclasis. In shear zones locally several inches wide the rock is crushed to a fine purplish-brown mylonite. Adjacent to the shear zones considerable areas of rock are brecciated; angular blocks of all sizes are embedded in a matrix of mylonite and crystal fragments. Except for minor development of chlorite in the mylonite, the crushed material shows no sign of recrystallization. Effects of crushing diminish with distance from the Colville batholith and on the hill west of the valley are represented only by occasional thin streaks of mylonite.

The malignites and other alkalic rocks of Kruger Mountain are discussed by Daly (1912) and more fully by Campbell (1939). Campbell finds no conclusive evidence on Kruger Mountain regarding the age of these rocks beyond the facts that they intrude Anarchist rocks and are intruded by the Similkameen granodiorite. East of Campbell's area the profound crushing of the malignite near the Colville contact indicates solidification before the intrusion of the Colville batholith, and the presence of a few small malignite xenoliths in the Osoyoos batholith establish the younger age of the latter.

INTRUSIVE ROCKS ROUGHLY CONTEMPORANEOUS WITH THE COLVILLE BATHOLITH

Osoyoos batholith.—Extending from British Columbia into the northern part of this area is the nose of a large granitic mass which Daly (1912) called the Osoyoos batholith. The mass apparently extends no farther south than the upper valley of Tonasket Creek, where it makes contact with the Colville batholith. To the north, across the border, the batholith widens considerably, but its northern and eastern limits are unknown. On the west it may include, as Daly believed, the mass of quartz diorite west of Osoyoos Lake.

Light-gray, fairly coarse-grained gneiss, with just sufficient orthoclase to make it a granodiorite rather than a quartz diorite, predominates in the batholith. A typical specimen differs from rocks of the Colville batholith in: (1) somewhat more abundant quartz and less abundant orthoclase (2) the nearly constant presence of epidote (3) the usual presence of hornblende and sphene (4) the lack of linear structure (5) characteristic platy aggregates of biotite, epidote, and hornblende. Nearly all Osoyoos specimens show cataclastic structures similar to those so pronounced in the Colville rocks—zones of finely granulated material, fragmented crystals, wavy extinctions. The crush zones show some recrystallization, and abundant patches and stringers of undeformed quartz and

orthoclase show that final solidification of the mass postdated the cataclasis.

In at least that part of the mass near the International Border structural and petrographic variations are much less pronounced than in the Colville batholith. Linear structure is inconspicuous except locally along the border, but platy structures defined by crush zones, by biotite flakes, and by biotite-epidote-hornblende aggregates are nearly always present. The attitude of the platy structure is fairly uniform, its strike averaging N. 45° W. and dips varying from 20 degrees to 90 degrees E. This attitude is similar to the dip and strike of the platy structure in adjacent parts of the Colville batholith and to the prevailing attitude of adjacent Anarchist rocks. In sharp contrast with the Colville batholith the Osoyoos mass is characterized by strong contact metamorphism and an abundance of diaschistic dikes along its borders. Fuller descriptions of the batholith are given by Daly (1912) and Krauskopf (1941).

Similkameen batholith.—The patches of Similkameen granodiorite in the west central part of the map (Pl. 1) are outliers of a huge mass lying just west of the mapped area. At the International Boundary, according to Daly, the batholith is 14 miles wide; its northern and southern limits are unknown, but it extends at least 40 miles south of the Border.

The Similkameen batholith is characterized by an almost complete lack of cataclastic structure and foliation. In some parts it is porphyritic, and along its borders dark colored variants occur rarely, but coarse-grained biotite-hornblende granodiorite, with only a moderate excess of plagioclase over potash feldspar and with quartz somewhat subordinate is the staple rock. Along its borders the batholith bakes Anarchist phyllites, quartzites, and conglomerates to hard black hornfelses. Both aschistic and diaschistic dikes occur near its contacts, locally in abundance.

Whisky Mountain stock.—A possible correlative of the Similkameen batholith is a porphyritic granodiorite exposed over about 6 square miles on Whisky Mountain. The stock is elongated roughly northwest-southeast, and apparently ends abruptly at the Okanogan Valley about 1½ miles from the nearest point of the Colville batholith.

Like the Similkameen rocks, specimens of this granodiorite show remarkable uniformity. Foliation and cataclastic structures are lacking except locally very near the contact. The quartz content is relatively small. In contrast to the rocks of the Similkameen batholith, Whisky Mountain rocks have a uniform abundance of microcline phenocrysts, an abundance of sphene and epidote, and no hornblende. Small green flakes of biotite often appear in aggregates with epidote; the aggregates frequently simulate the shapes of amphibole crystals. Contact meta-

morphism near the stock is limited to mild baking. Aschistic dikes and quartz veins are abundant in the Anarchist rocks for some distance from the intrusive contact.

Chelan batholith.—The Chelan Mountains, lying about 60 miles south of the map area, are underlain by a very large igneous mass named the Chelan batholith (Waters, 1938). This mass extends an unknown distance northwest and north, and to the east it disappears beneath the Columbia River lavas. Just south of the map area the wedge-shaped mass of Anarchist rocks between the Similkameen and Colville batholiths disappears, and the two intrusives come into contact. Still farther south, perhaps, these two igneous bodies make contact with the Chelan mass. However, it is possible that in the unmapped region to the south the three batholiths may merge instead of showing clean-cut contacts.

Where the Chelan batholith has been studied in detail it corresponds rather closely to the Colville mass in composition but not in its structural features. Although flow structures and protoclastic effects are locally developed near the margin of the Chelan mass they are not conspicuous, and most of the batholith has no uniformly oriented structures. The batholith consists mainly of biotite-hornblende granodiorite with local quartz monzonite and quartz diorite variants. Biotite is usually the predominant ferromagnesian mineral, but diminishes or disappears where the granodiorite makes contact with hornblende schists. The latter contact zone is further characterized by innumerable blocks of hornblende schist rifted off the wall and in all stages of recrystallization, mechanical shattering, and differential assimilation. This remarkable contact breccia zone is locally as much as 4 miles thick. In composition the resemblance of this zone of contact breccia to the migmatitic facies of the Colville batholith is striking. However, the two differ markedly in structure. The border zone of the Chelan mass is an arteritic migmatite. It does not have the streaky banding nor the cataclastic textures characteristic of the border of the Colville mass.

ROCKS YOUNGER THAN COLVILLE BATHOLITH

Spilite and keratophyre.—In the central part of the area are two exposures of somewhat altered volcanic rocks which unconformably overlie Anarchist strata and are themselves overlain by Tertiary andesites and conglomerates. Mostly massive lavas, they include some volcanic breccias. Their maximum thickness is at least 150 feet. From Anarchist greenstones the lavas are distinguished by the excellent preservation of vesicles, flow structures, and phenocrysts; by lack of intense shearing; and by the complete absence of epidote. From the later Tertiary lavas they are set off by their prevailing green color, the decomposition of their mafic

minerals to chlorite, the presence of albite as their only feldspar, and local minor zones of shear. The rocks are cut by none of the dikes and quartz veins common in the surrounding Anarchist beds. No boulders from the lavas were recognized in any of the early Tertiary conglomerates. Such evidence does not fix the age of the lavas definitely but makes their assignment to the early Tertiary at least probable.

The least altered specimens have feldspar and/or pyroxene phenocrysts in a pilotaxitic groundmass of the same two minerals. Always a part and often all of the pyroxene is altered to chlorite and calcite. The feldspar, in phenocrysts and in the groundmass, is entirely albite (An_{07} to An_{02}). The majority of specimens correspond in mineral composition to spilites, although some have sufficient feldspar to be called keratophyres. The single available chemical analysis shows a silica content of 49.2 per cent and a soda content of 4.3 per cent (calculated on a H_2O-free and CO_2-free basis).

Sedimentary rocks.—Clastic Tertiary sedimentary rocks occur in three parts of this area: on both sides of the Okanogan Valley and for several miles up the Similkameen Valley near Oroville; on Whitestone Mountain and adjacent hills across the Okanogan Valley; and in the valley of Pine Creek. Sandstones and conglomerates are most abundant, siltstones and shales appearing only locally. The angular grains, poor sorting, and irregular bedding of the coarser rocks suggest deposition by rapid mountain streams; some of the fine shales may have been deposited in temporary lake basins. Plant fossils preserved in the shales fix the age as early Tertiary, most probably Oligocene.

The thickest Tertiary section is exposed on the south wall of the Similkameen Valley, where 2000 feet is visible between the unconformable contact with the Anarchist series at the river and the top of the hill. Sandstones and coarse boulder conglomerates alternate erratically. The conglomerates are particularly interesting. Their fragments are subround to subangular, most of them ranging from a fraction of an inch to a foot or so in diameter, a few up to 6 feet. Granodiorite similar to that in the Similkameen batholith is by far the most abundant rock; Anarchist fragments are abundant locally, and near the river boulders of the alkalic rocks of Kruger Mountain become conspicuous. The matrix is gray-green to buff and is often in part very fine-grained. Frequently the conglomerates show no sign of bedding except for occasional thin sandy layers.

On Whitestone Mountain brilliant white beds of coarse sandstone with abundant tuff fragments crop out. Coarse conglomerate is also common here, some full of large tuff boulders. East of the Okanogan Valley is a

conglomerate overlying andesite and containing abundant andesite fragments.

In the valley of Pine Creek the Tertiary rocks are principally sandstones and siltstones, with a thin basal conglomerate containing abundant tuff as well as Anarchist and Similkameen fragments.

Structures in the Tertiary beds consist chiefly of broad, gentle folds with a prevalence of eastward dips. Clearest of the folds is a broad syncline trending northwest-southeast in the thick Tertiary section just south of the Similkameen River. In one small locality in the valley of Tonasket Creek the Tertiary beds are vertical, but elsewhere no dips over 40 degrees were noted.

Andesite and dacite.—The Tertiary andesites and dacites are a variegated assortment, most of them showing the effects of slight hydrothermal alteration. They consist of flows, tuffs, breccias, and volcanic necks, among which no distinction is made on the map (Pl. 1). Andesites crop out in three principal areas: (1) near Oroville; (2) on Whitestone Mountain and adjacent hills; and (3) east of Pine Creek, where dacite flows also occur. The greatest exposed thickness of andesitic material, estimated at 2000 feet, is east of Pine Creek; the greatest observed thickness of a single flow is 360 feet. The bulk of the volcanic material is restricted to the upper portion of the Tertiary section. Megascopically the andesites are porphyritic-aphanitic rocks, with hornblende and andesine the most common phenocrysts. Gray is the usual color, but shades of purple, green, and brown are not infrequent. Platy jointing is often conspicuous, and flow banding is well developed locally. More altered varieties have lighter colors and a characteristic chalky appearance. The andesite tuffs are usually brilliant white or light buff.

Microscopically the fresher andesites show a hyalopilitic or pilotaxitic groundmass made up of andesine, augite, hornblende, and glass. In many specimens hydrothermal alteration has introduced pyrite and converted much of both groundmass and phenocrysts into chlorite, calcite, and sericite. A curious feature of the alteration locally is the partial or nearly complete replacement of andesine by albite, the albite invading along cracks and cleavage planes.

At the east end of Kruger Mountain is a large volcanic neck of massive andesite and at least two smaller satellitic necks cutting Tertiary conglomerates. Several volcanic necks can also be recognized in the Pine Creek area. Each of the three isolated masses of andesite located in Mud Lake Valley (Pl. 1) is a neck which has drilled up through the Middle Anarchist series. The rock in the largest neck has been thoroughly decomposed by hydrothermal alteration. Other necks with accompanying dikes

may be found in the large andesite area east of Pine Creek. Elsewhere in the area, except for a few small dikes, outcrops consist of erosional remnants of flows and pyroclastics.

The dacitic flow rocks of the Pine Creek area constitute only an insignificant part of the entire volcanic succession, though they may be more abundantly represented in the pyroclastics. The most conspicuous flow, in most localities over 200 feet thick, occurs in the upper part of the section. This flow is a normal biotite dacite.

Glacial debris.—Fresh glacial boulders near the highest summits prove that this area was completely covered by ice during the last Pleistocene glaciation. Retreat of the glacier left the lower parts of the region mantled with a thick cover of glacial debris. Persistence of ice in the Okanogan Valley long after it had disappeared from the higher slopes led to deepening of the valley and an especially thick accumulation of debris, mainly deposited by meltwater. None of the observed glacial material could be definitely ascribed to glaciation older than the latest stage, though intricately deformed glacial lake silts which probably owe their structure to overriding by an ice mass are conspicuous from Tonasket southward, and are unconformably overlain by till south of the map area near Omak.

Till is limited to thin veneers on the gentler slopes and to small morainal ridges especially well developed in the upper valley of Ninemile Creek and in the valley of Antoine Creek. Most of the glacial material has been stratified by water and deposited as terraces, some lacustrine, but the majority stream terraces built at the sides of melting ice tongues.

The stream terraces are exceptionally well preserved. Large ones flank the Okanogan River, and smaller ones extend in unbroken succession up tributary valleys. They are most prominent in the valley of Ninemile Creek, less conspicuous in the valleys of Antoine, Bonaparte, Tunk, and Aeneas creeks. The highest ones reach an elevation of about 3600 feet.

The smaller high level terraces consist of cross-bedded sand, gravel, and silt. The broad terraces along the Okanogan River have a similar composition locally, but more commonly consist chiefly of silt with a gravel veneer. Surfaces of these low level terraces are pitted with kettles, some up to 1000 feet in diameter. Movement of ice against the large terraces as they were forming is recorded by local intense crumpling of the silt beds.

Broad, level terraces, probably formed in ice-dammed or debris-dammed lakes occur in several parts of the area. The most prominent ones are near Osoyoos Lake, and northwest of Whitestone Mountain. The rela-

tions of similar glacial terraces at localities farther south in the Okanogan and Columbia valleys have been described in detail by Waters (1933), and Flint (1935) has also given a general account of glaciation in the Okanogan Valley.

THE COLVILLE BATHOLITH

RELATION OF THE BATHOLITH TO ITS WALLS

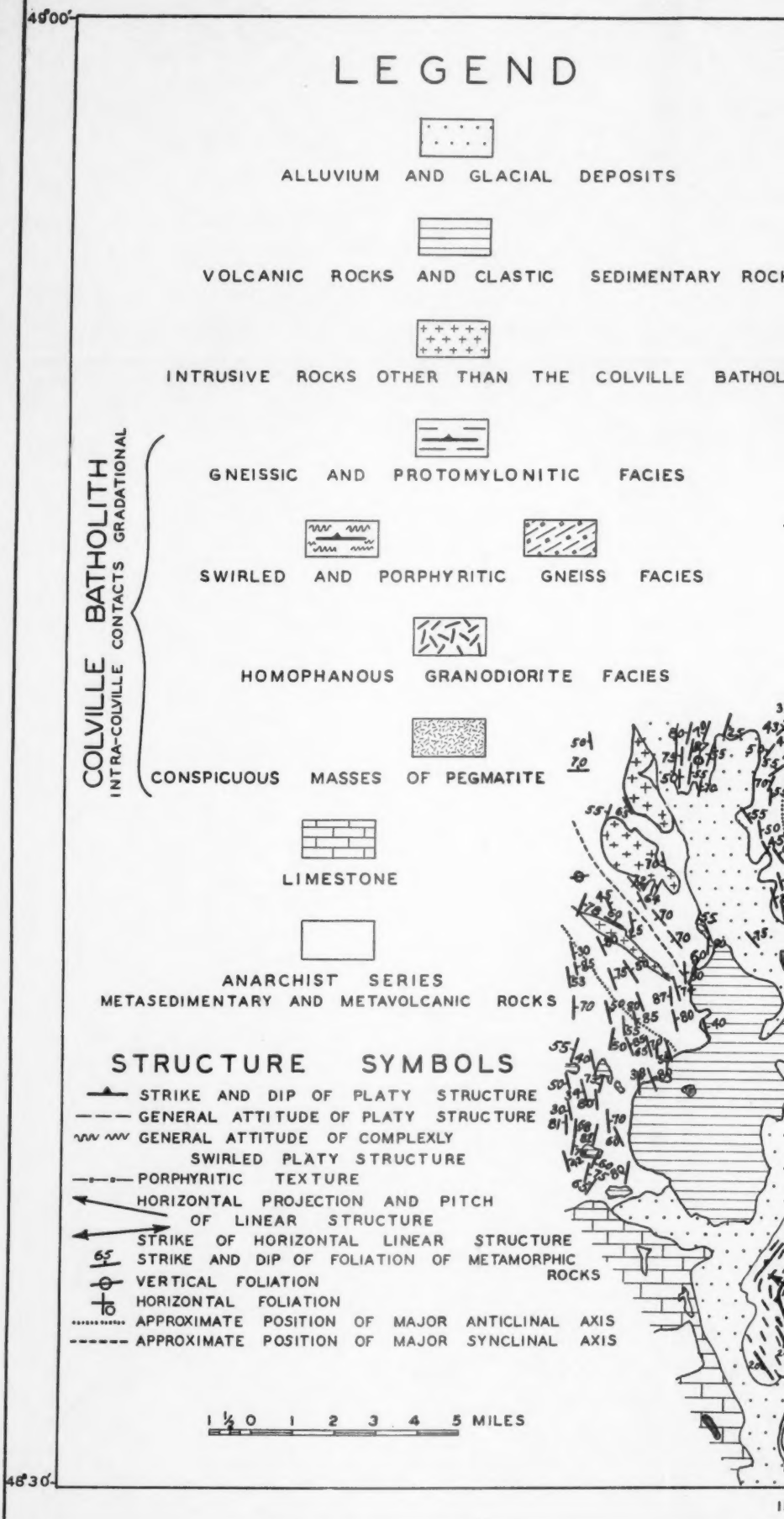
Discordant character of contacts.—Into the heterogeneous Anarchist series and associated intrusive rocks the Colville batholith was intruded. The trend of the folds in the Anarchist as well as the direction of foliation and other structural features is predominantly northwest-southeast, and across these structural trends the batholithic border cuts with decided discordance. Folds, faults, elongate igneous intrusions, foliation, and bedding are all transected at an angle (Pls. 1, 2). Only locally is there any suggestion of concordance between the batholithic contact and earlier structural trends.

Zone of brecciation at periphery.—Despite the abruptly transgressive contact, however, the Colville mass cannot be described as a "stoping batholith." The walls of the intrusive are not veined by apophyses, nor are disengaged angular blocks of wall rock common in the intrusive. Smeared, pulled out, partially assimilated inclusions are found in abundance, but differ from the immediately adjacent rocks beyond the contact. The absence of angular inclusions of the adjacent wall rock in the border portion of the batholith is all the more striking considering the nature of the wall rocks for some distance out from the contact. Usually for a quarter of a mile or more, the rocks bordering the batholith have been made into a gigantic friction breccia. Small faults penetrate everywhere, and along them movement has occurred resulting in the rotation of individual blocks with respect to one another so that the foliation and bedding near the contact are at all angles. Yet no great displacement of the individual blocks has occurred, for in general a band of greenstone in the Anarchist series that trends into the contact will have a greenstone breccia at its end, and limestone masses end similarly in highly brecciated and mashed limestones which contain few foreign fragments. Although the breccia fragments vary greatly in size they are generally large. Fragments less than a few feet in diameter are rare except in the zones of comminution between adjacent blocks, and many blocks have dimensions of 100 feet or more. Unfortunately, the exposures of this breccia zone are not very good; for, being readily eroded, this zone has largely determined the course of the Okanogan River and consequently is covered in many places by alluvium and glacial debris. However, remnants are preserved at many places along both walls of the valley.

Within the breccia itself the small faults and zones of shear bounding the individual blocks are so chaotic in trend and dip as to make structural mapping almost impossible. Nevertheless, despite innumerable joints and faults with every possible trend and dip, three directions of fracture are somewhat better developed than others. One set strikes roughly parallel to the contact, and dips into the batholith at angles of 10 to 50 degrees. This set is generally slickensided, indicating thrusting upward and away from the intrusive. Another set has the same strike, but the dip is very steep or vertical, and vertical slickensides appear on the fault planes. The third common set usually trends approximately at right angles to the contact, though locally deviating as much as 45 degrees, dips vertically, and may or may not show slickensides. If slickenside striae are present they usually pitch at a low angle toward the intrusive or else are horizontal. It should be emphasized that in addition to these three sets chaotic joint assemblages trend in all directions, but even these joints show some semblance of order, for when the pitch of the striae on slickensides is plotted a general motion of the intrusive upward and outward with respect to its walls is indicated. For a given area along the contact, the projected strike of striae on slickensides commonly deviates little in direction irrespective of the strike and dip of the slickensided joint plane. However, the pitch of the striae may vary from horizontal to vertical. Many local deviations may possibly be explained by rotation of blocks after the slickensides were made.

Locally resistant masses of wall rock have not been broken down into a friction breccia. The strong gabbros on Earl Mountain are broken into large fragments but are not thoroughly brecciated. On the other hand, a mass of malignite north of Tonasket Creek has been so thoroughly crushed that portions resemble a conglomerate. The massive greenstones of the Middle Anarchist on the ridge east of Antoine Valley are not notably sheared. However, these greenstones are not at the walls of the batholith, but form a part of its roof. Although brecciation is profound along the walls, it appears to be less conspicuous in the batholithic cover.

Absence of contact metamorphic effects.—The wall rocks at the periphery of the Colville batholith show almost no evidence of contact metamorphism, particularly where the gneissic and mylonitic facies of the batholith meets the wall rocks. There is little or no evidence of baking, or of the additive contact metamorphic effects that commonly occur along the borders of large batholiths. Even such susceptible rocks as impure limestone and chlorite-rich greenstone are completely unchanged. In fact, throughout wide areas along the contact, the only metamorphic effects that can be ascribed to the batholith are the development of films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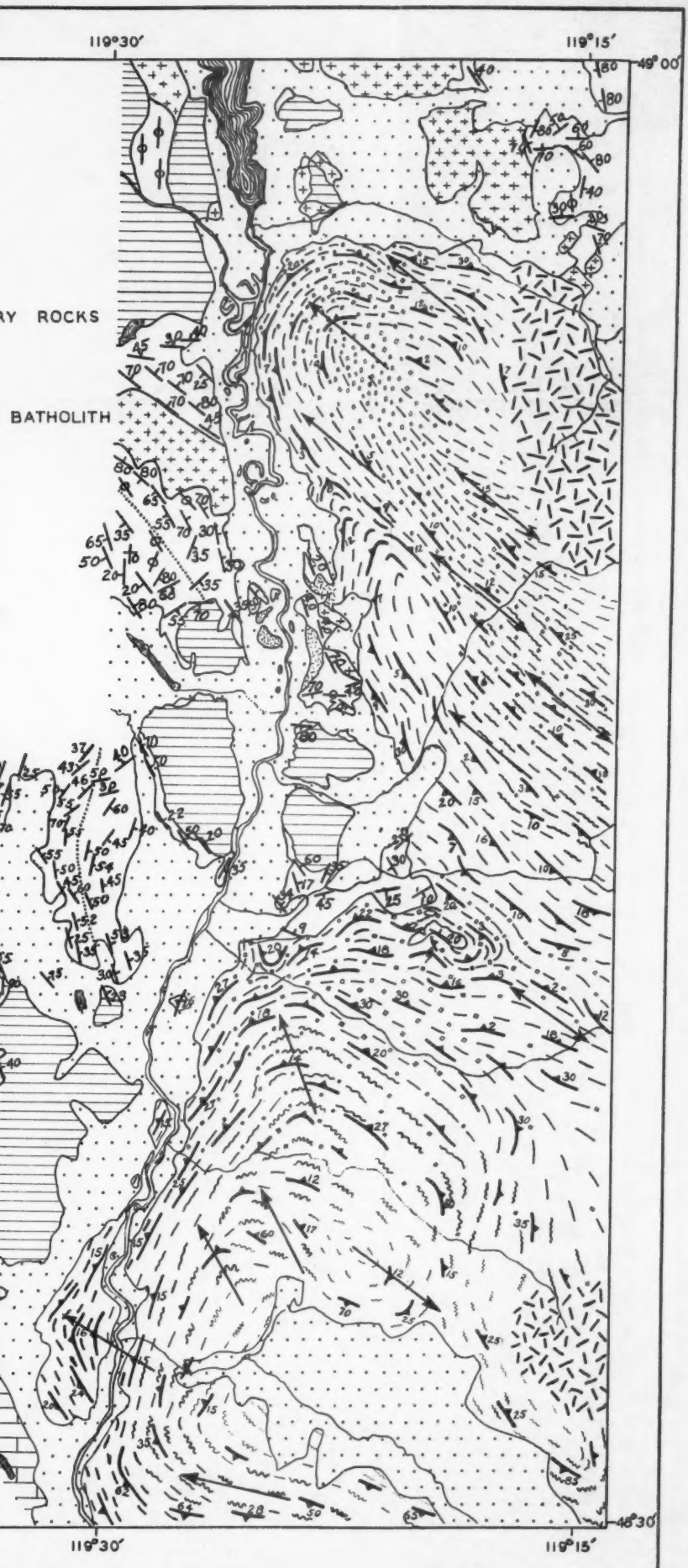
LEGEND
ALLUVIUM AND GLACIAL DEPOSITS
VOLCANIC ROCKS AND CLASTIC SEDIMENTARY ROCK
INTRUSIVE ROCKS OTHER THAN THE COLVILLE BATHOL
COLVILLE BATHOLITH
INTRA-COLVILLE CONTACTS GRADATIONAL
GNEISSIC AND PROTOMYLONITIC FACIES
SWIRLED AND PORPHYRITIC GNEISS FACIES
HOMOPHANOUS GRANODIORITE FACIES
CONSPICUOUS MASSES OF PEGMATITE
LIMESTONE
ANARCHIST SERIES
METASEDIMENTARY AND METAVOLCANIC ROCKS
STRUCTURE SYMBOLS
STRIKE AND DIP OF PLATY STRUCTURE
GENERAL ATTITUDE OF PLATY STRUCTURE
GENERAL ATTITUDE OF COMPLEXLY SWIRLED PLATY STRUCTURE
PORPHYRITIC TEXTURE
HORIZONTAL PROJECTION AND PITCH OF LINEAR STRUCTURE
STRIKE OF HORIZONTAL LINEAR STRUCTURE
STRIKE AND DIP OF FOLIATION OF METAMORPHIC ROCKS
VERTICAL FOLIATION
HORIZONTAL FOLIATION
APPROXIMATE POSITION OF MAJOR ANTICLINAL AXIS
APPROXIMATE POSITION OF MAJOR SYNCLINAL AXIS
1½ 0 1 2 3 4 5 MILES
48°30'

STRUCTURAL MAP OF PART OF BORDER OF



RDER OF COLVILLE BATHOLITH

and crusts of epidote and quartz along the faults and joints that subdivide the fragments within the zone of friction breccia.

In a few places, however, where other facies of the batholith make contact with the wall rocks, contact metamorphic effects are more marked. Pegmatites are abundant near the border and locally bake the wall rocks to a hornfels, but metasomatic changes are more common. The transformation of the gabbro on Earl Mountain as a result of emanations from one of these pegmatite bodies has already been mentioned.

Locally the swirled and porphyritic gneiss facies of the batholith is found at the contact or else is separated from the wall by only a thin shell of gneissic or mylonitic rocks. Here, the metamorphic effects are greater, though not profound. The main change is an increase of epidote and mild sericitization and propylitization of the wall rocks.

Near Evans Lake, just southwest of the mapped area, an igneous mass, possibly an offshoot of the Colville, is composed of equivalents to the porphyritic gneiss facies and the homophanous [1] facies of the main batholith. Adjacent to this offshoot contact metamorphic effects are really great. Impure limestones have been changed for considerable distances into coarse-grained lime silicate rocks; phyllites and slates become chiastolite-mica schists, and quartzite bands in the wall rocks show drastic recrystallization.

Thus it appears that the Colville mass locally effected strong metamorphic changes, but in the majority of places the gneissic and mylonitic zone along its border formed a protective blanket which held in the heat from the interior of the mass and allowed no metamorphic changes except the profound brecciation of the wall rocks, and local veining by epidote and quartz. In general, many of the characteristics of this contact suggest a fault contact rather than an intrusive contact. From detailed consideration of relations both within the wall rock and within the adjacent intrusion, however, it is believed that the features of the contact zone can best be attributed to the rise and emplacement of the batholith as a unit. By this interpretation the intense brecciation and usual lack of metamorphism in the wall rocks are due to a rise of the intrusive at a time when its peripheral portion (which now forms the gneissic and mylonitic facies) was already nearly solid.

The relations described above hold only along the walls of the batholith. In those rare places where pendants of rock from the roof have survived erosion the contact effects are much greater. One of the most interesting places to view this relation is along the lower part of Siwash Creek and on the divide between Siwash Creek and Bonaparte Creek. Here green-

[1] "Homophanous" means "without directional structure" Sederholm, (1923, p. 73).

stones have been veined by the intrusive and converted into migmatitic rocks. Porphyroblasts of microcline have developed in abundance in the shattered and veined greenstone bands and the matrix of the greenstone has been coarsened by recrystallization until it resembles a plutonic rock. However, even here the effects of migmatization are much obscured by the profound mashing and granulation which has affected all rocks included in the gneissic and mylonitic zone along the border of the batholith.

Absence of satellitic dikes in the wall rocks.—The contact zone around the border of the Colville batholith is unusual for the rarity of satellitic dikes. In this respect the batholith stands in marked contrast to the Chelan, Osoyoos, and Similkameen batholiths, and to the near-by Whisky Mountain stock, all of which are characterized by abundant apophyses and by diaschistic dikes. Pegmatite and alaskite masses from the Colville batholith locally penetrate the wall rocks in great volume. However, these seldom occur in dikes or other injected masses of definite shape but appear to have oozed out of the peripheral portion of the batholith and to have formed irregular-shaped bodies at the contact or else to lie as podlike masses just within the batholithic border. The pegmatites are commonly abundant where the contact of the mass turns abruptly, as for example along the batholithic border north and northwest of Mount Hull.[2] The gneissic and mylonitic border of the batholith is not diked by rocks of its own composition, nor does it send apophyses into the wall rocks. In fact the writers have observed no aschistic dikes anywhere in the Colville mass. Lamprophyre dikes are found locally near the summit and north of Mount Hull, and very rarely in other parts of the Colville mass, but the writers have not observed them cutting the wall rocks outside the contact.

STRUCTURAL FACIES OF THE BATHOLITH

General statement.—For purposes of structural mapping the heterogeneous rocks of the Colville batholith have been divided into three main groups. The first comprises prominently banded rocks that occur at the periphery of the mass in a zone ranging from only a few yards up to 4 miles in thickness. Petrographically this zone is more heterogeneous than any of the others and commonly is made up of alternating dark and light layers. Because of the widespread development of mylonites and gneisses of various kinds within it, it has been designated the "gneissic and mylonitic facies." This zone grades inward into a second broad zone of less conspicuously foliated rock in which the planes of foliation have commonly been thrown into a series of swirls and minor contortions. Else-

[2] Following local usage Mount Hull in this report refers to the triangular-shaped tract of mountainous land lying east of Okanogan River, southwest of Tonasket Creek, and north of Mosquito Creek, and does not refer specifically to the small eminence on which the lookout is located.

FIGURE 1. NEAR RIVERSIDE
Gneissic and mylonitic facies of batholith forms mountains on skyline.

FIGURE 2. NEAR MOUTH OF TUNK CREEK
Rocks of gneissic and mylonitic zone at left. Rounded hill at right underlain by swirled gneiss.

FIGURE 3. SOUTHWEST OF MOUNT HULL
Shows banding of light and dark material, conspicuous foliation, and two major joint sets in gneissic and mylonitic zone.

MARGIN OF COLVILLE BATHOLITH

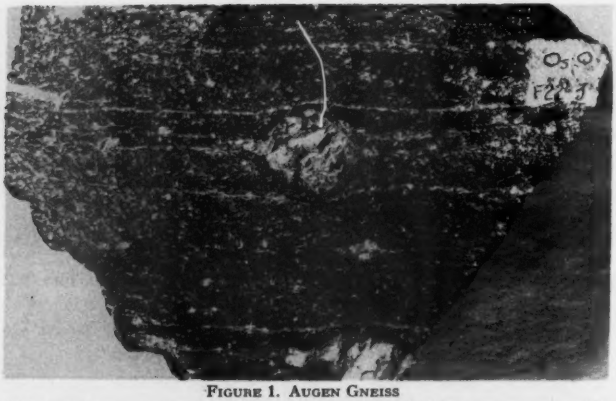


FIGURE 1. AUGEN GNEISS
McLaughlin Canyon. Porphyroclasts of microcline, crushed into augen, surrounded by blastomylonite matrix.

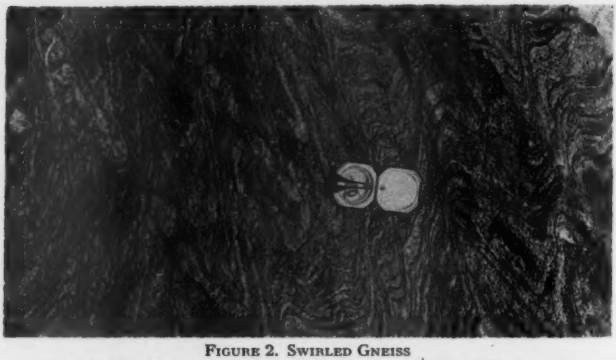


FIGURE 2. SWIRLED GNEISS
Tunk Creek Valley. Note shears which cut the contorted folia on right.

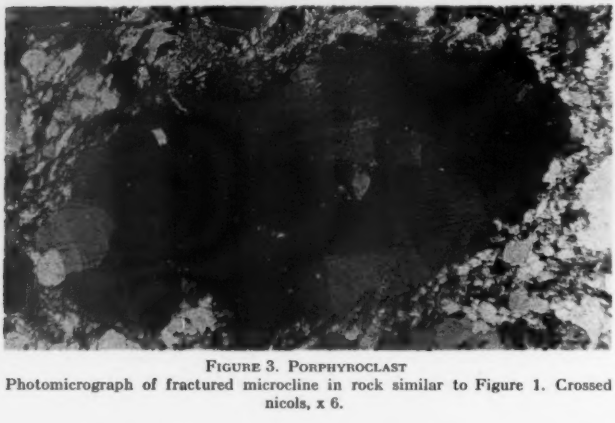


FIGURE 3. PORPHYROCLAST
Photomicrograph of fractured microcline in rock similar to Figure 1. Crossed nicols, x 6.

ROCKS FROM BORDER OF COLVILLE BATHOLITH

where the foliation is less wavy, and the rock has a marked porphyritic texture. Because of these characters this zone has been designated the "swirled and porphyritic gneiss facies." It is generally much more uniform petrographically than the gneissic and mylonitic facies. In the Colville Range the swirled and porphyritic gneiss facies gives way gradually to homophanous igneous rock forming the cores of four conspicuous mountains: mounts Hull, Bonaparte, Annie, and Tunk. Only Mount Tunk and Mount Hull lie within the map area, but Mount Bonaparte and Mount Annie are only a short distance east. The uniform massive plutonic rocks of these four centers have been designated the "homophanous facies" of the Colville batholith.

Between these three facies very gradual transitions occur. The lines drawn on Plate 1 separating them are arbitrarily chosen and are not in any case to be interpreted as marking sharp contacts.

Gneissic and mylonitic facies.—The most conspicuous feature of the rocks in this zone is their pronounced foliation. Viewed from a short distance they resemble a series of thin-bedded sediments (Pl. 3, figs. 2, 3) and the rifting along the foliation planes together with erosion along a prominent series of joints which cuts the foliation at right angles, is responsible for many scenic features of the district (Pl. 3, fig. 1).

In outcrops near the contact with the older wall rocks the material of this zone is usually a medium gray though locally it is quite black. At most localities the dark surface is broken by numerous white spots which locally may be so small as to be scarcely visible, or may reach a diameter of as much as 1½ inches (Pl. 4, fig. 1). These are crystals of potash feldspar whose edges have been crushed until the resultant spot is augen-shaped and frequently has trails of crushed feldspar bits extending out from either end. The evidence that these crystals have been crushed by rolling within the foliation planes is clear, and at many localities the direction of rolling can be told by the deviation of foliation planes. Where the direction of movement can be determined it indicates that the central portion of the batholith was moving upward with respect to a more stationary wall.

The matrix between the crushed microcline phenocrysts is variable. In specimens taken very close to the contact it is usually dark gray to black, and the microscope reveals minutely crushed and granulated mineral grains and the microstructure of a mylonite. Locally near the border some parts of the matrix are bright black, lustrous, and unfoliated. These lustrous bands generally parallel the foliation planes but in a few cases cross them at an angle. Viewed microscopically these bands consist of a nearly opaque isotropic base highly charged with fragments of crushed

minerals. This material resembles rocks described as pseudotachylyte and ultramylonite (Shand, 1916; Waters and Campbell, 1935) in other parts of the world.

In the zone immediately adjacent to the contact the foliation trends parallel to the wall, and the dip of the foliation also approximates the dip of the contact very closely. The foliation planes are in general remarkably smooth and straight, but when viewed in cross section on slabs several square feet in area the foliation planes are seen to thin and lens out, to be affected by minute rolls and waves, and even apparently to criss-cross each other at extremely acute angles. In other words the foliation in many ways bears a more marked resemblance to the flow banding of a highly laminated dacite or rhyolite than it does to a truly metamorphic rock.

Within the foliation planes lineations of various kinds are sometimes apparent. These are elusive and have not been studied carefully enough to justify any very definite statements about them, but certainly more than one kind of linear structure is present, though perhaps all lineations have been produced during the same period of movement and do not represent successive epochs of deformation. One kind of lineation is furnished by the previously described feldspar augen. At some localities the feldspar augen clearly have been rolled into spindle shapes, and in many places augen tend to cluster within the foliation plane. The rodding of individual augen is not marked, however, and the longer axis is seldom more than twice the length of the shorter axis. The augen always lie within the plane of foliation and at several places where they have been measured they lie horizontal or pitch very slightly. They have not been sufficiently investigated in other areas to warrant the statement that horizontally elongated augen surround the batholith at its border, but the local evidence suggests it.

Another type of alignment commonly encountered is the presence of minute striae on slickensided planes of foliation. These striae parallel or nearly parallel the dip of the foliation. Their projected strike is therefore at right angles to that of the rodded feldspar augen, yet they have doubtless been produced by the same movements, *i. e.*, the augen by rolling within the planes of foliation, the striae by slip movement of one foliation plane against another.

Other striae appear on slickensided surfaces which do not correspond with foliation planes. The projected strike of these is often the same as that of the striae on slickensided foliation planes, but their pitch is commonly widely divergent. In general they follow the same pattern as the striae previously described as occurring in the zone of brecciated country rock beyond the molar contact, thus confirming the conclusion that they

were produced as a result of movement upward and outward from the central portion of the batholith.

Some distance in from the contact the mylonitic paste that fills the space between the feldspar augen has been largely recrystallized. In this newly crystalline matrix lineation is sometimes shown by individual elongate minerals. The relations of this kind of lineation are complex and have not been fully investigated. In addition, within this more recrystallized zone inclusions and swarms of inclusions are found locally which have been pulled into linear structures, and furthermore the foliation planes, particularly as the zone of swirled and porphyritic gneisses is approached, are commonly wrinkled and contorted.

The trend of the clots of inclusions or of aggregates of dark minerals such as biotite flakes and of the axes of plications of the foliation planes forms the least elusive and apparently most consistent linear structure, and this structure has been plotted on the structural map (Pl. 2). Its projected strike is invariably northwest-southeast, parallel to the major structural grain of the Anarchist rocks. This linear structure is also conspicuously developed in the swirled and porphyritic gneiss facies.

Many other interesting minor structural features occur within the zone designated as the gneissic and mylonitic facies. Much of this zone, prior to the movements which have given it its remarkable foliation, appear to have been in the condition of a migmatite. The country rock was a mafic type containing abundant amphibole and was split by the invading magma into slablike inclusions, probably along a direction of earlier schistosity. This mixed border zone has taken part in the mechanical movements and many structural details have resulted because of the difference in competency between host and partially crystallized intrusive in the migmatitic zone. The slablike inclusions of hornblendic rock have been swung around and oriented within the foliation planes. Furthermore, they show the effects of pulling out by the matrix enclosing them. Many are lenses with blunt terminations. One inclusion may be severed several times and the ends rounded off so that a series of podlike bodies of amphibole-rich rock lie enclosed in a mylonitic matrix. In some cases the inclusion was sufficiently brittle so that the broken ends did not round off, and in still other cases the breakage resulted in the production of excellent examples of boudinage structure (Wegmann, 1932). Occasionally the gneissic matrix between adjacent inclusions has been deformed in incompetent fashion between the more competent hornblendic rock on either side resulting in markel drag folding of the matrix between the bands (Fig. 2). Consistent with other data, the orientation of the axial planes of the drag folds indicates that the inclusion toward the center of the batholith was rising with respect to a more stationary wall.

Additional evidence of this movement is locally afforded by thin pegmatite dikes which cross the direction of foliation. These dikes, though fairly well defined, have borders in which the mineral grains lock across the contact with the mineral grains of the enclosing rock, and they appear

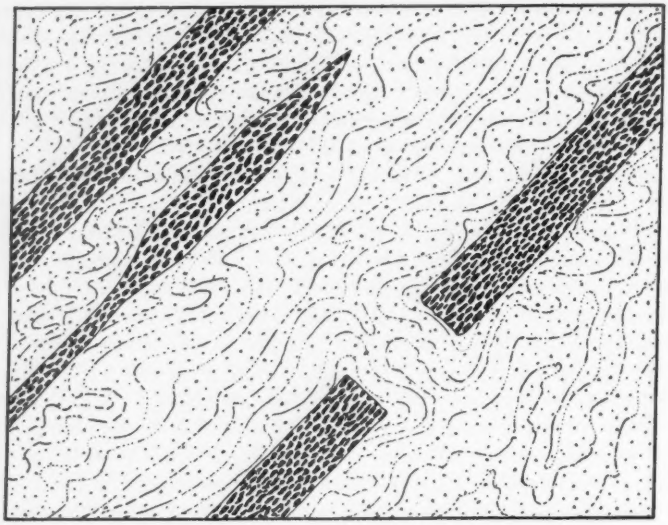


FIGURE 2.—*Cross section of drag folds*

In protoclastic gneiss between recrystallized hornblendic inclusions. Batholithic contact toward left. Length of section 2 feet. Bonaparte Creek.

to have been formed by the oozing of a residual mass of liquid material out of the gneissic and protomylonitic facies into fractures that were formed by deformation near the closing stages of solidification. However, the interesting feature about these dikes is that although they appear to form a fairly straight line across a cliff from a distance, their course when traced in detail is very sinuous and irregular. They will cross a few folia at right angles, then bend parallel to the foliation plane for a short distance, and again cross a few more folia; this is repeated many times in a few feet. In short, their course suggests a miniature stairway save that the sharp corners where treads and risers meet have been rounded off, or sometimes made into bulblike expansions. The explanation suggested for these interesting dikes is that they developed as cross joints (in the Cloos terminology) during the closing stages of movement in the gneissic and mylonitic zone. They were simultaneously filled

with pegmatitic liquors squeezed out of the adjacent rock. However, movement on the foliation planes had not yet entirely ceased, and as it continued the channelway of the still fluid dikelet was offset repeatedly by foliation plane slip. Direction of offsetting shows that the lower inte-

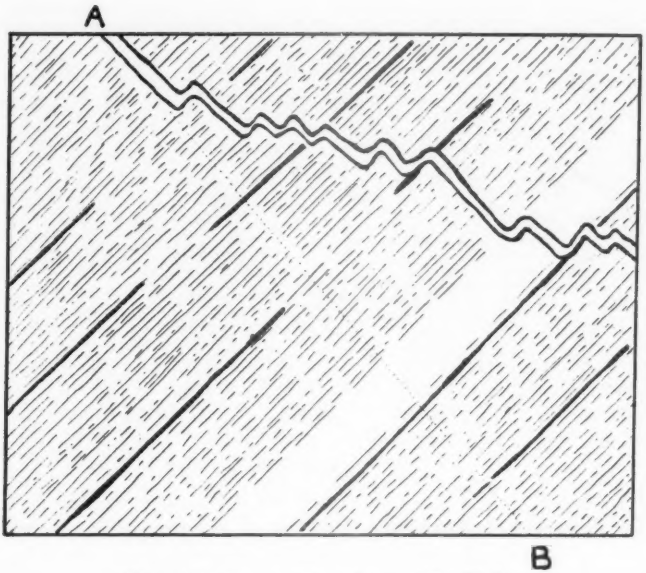


FIGURE 3.—*Cross section of pegmatite dikelet*
Repeatedly offset by foliation plane slip. Original fracture indicated by A-B. Batholithic contact to left. Natural size.

rior part of the dikelet was dragged upward with respect to the more stationary outer portion of the batholith. In this way the dike achieved its sinuous stairlike outline, and at the same time acquired a general dip no longer at right angles to the dip of the foliation and of less amount than at the time the original fracture formed (Fig. 3).

Swirled and porphyritic gneiss facies.—The outer zone of the Colville batholith with its numerous structural complexities passes gradually into an inner zone which though decidedly heterogeneous, is much more uniform both structurally and petrographically than the gneissic and mylonitic facies. This zone, which has been designated the swirled and porphyritic gneiss facies, is composed mainly of gneissic granodiorite. Locally it is markedly porphyritic and contains abundant phenocrysts of microcline.

Foliation in this zone is well-marked but not as conspicuous as in the previously described rocks. Furthermore, the foliation over considerable areas is not straight, but passes through the most involved convolutions and swirls (Pl. 4, fig. 2). The swirling in some areas is

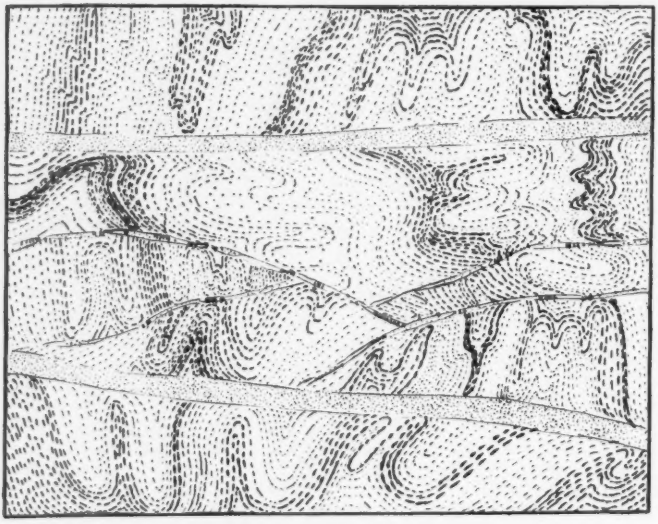


FIGURE 4.—*Swirled gneiss*

Contorted flow planes cut by shears (center of figure) and by zones of recrystallized sheared material (stippled areas). Drawing represents 6 square feet.

so highly complicated and appears so thoroughly folded in all three dimensions that the term ptygmatic is correctly applied to it; elsewhere the swirling may be less involved and definite axes of plication can be determined and plotted on maps. At some localities the swirling is on a large scale, the individual plications measuring several feet or even tens of feet across. More commonly the amplitude of folds is several inches or a few feet. Entire areas of several square yards may be uniformly swirled, or in other cases the swirled areas may lie in bands parallel with the foliation in the outer zone of gneissic and mylonitic rocks. Such bands are usually separated either by thin and sometimes discontinuous zones of gneisses similar to those making up the outer zone of gneisses and mylonites, or more commonly by a medium- to fine-grained gray rock with a granulitic texture which appears to be a thoroughly recrystallized mylonite (Fig. 4). Elsewhere contorted bands

are separated by complicated zones of shear (Fig. 5) and the granulitic rock appears to be material which has recrystallized along similar shear zones. The alternation of zones of swirled material with material of gneissic or granulitic character is particularly common in areas where

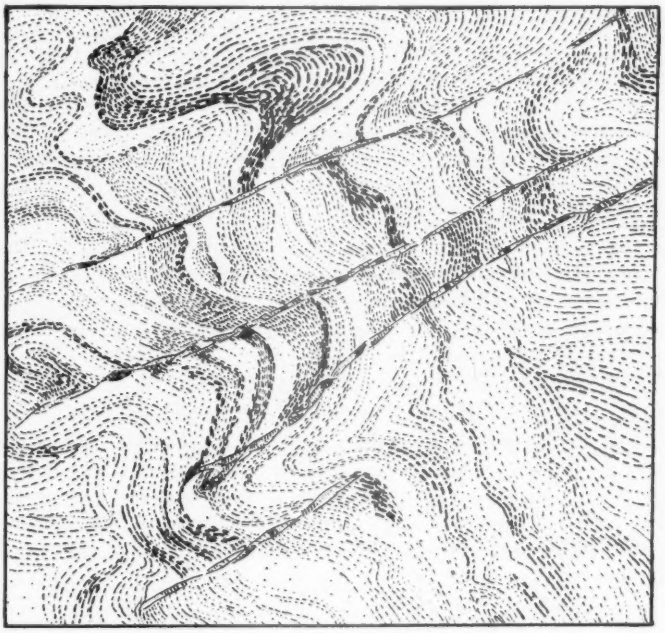


FIGURE 5.—*Swirled gneiss*

Ptygmatically folded flow planes cut by nearly contemporaneous shears. Drawing represents 16 square feet of outcrop.

the transition between the gneissic border and the swirled interior occurs. In fact the transition is accomplished simply by the appearance of swirling in certain bands, and the gradual increase in the number of swirled bands until practically the entire rock shows the contorted character.

However, not all parts of this inner zone are characterized by swirling. In some places the foliation is not contorted but is straight and continuous over large areas. Here the distinction between this zone and the outer gneissic and mylonitic zone is based largely on petrographic criteria—the inner zone is more homogeneous; under the microscope it shows the texture of an igneous rock; and cataclasis is conspicuous

only in the phenocrysts, the groundmass being relatively free of strain. Furthermore, the foliation, though plainly marked, is much less conspicuous than in the outer zone. The porphyritic gneisses above described grade into the gneisses of the border zone by gradual increase of the features due to cataclasis and recrystallization, and corresponding increase in the intensity of foliation.

Within the zone of swirled and porphyritic gneisses, two structural features are sufficiently conspicuous to be readily plotted. The most obvious one is again the strike and dip of the foliation. The attitude of the foliation on a regional scale conforms very well with that of the foliation in the outer zone of gneisses and mylonites, though in general the dip is less steep. Throughout the Colville batholith the dip decreases away from the contact. Errors may easily be made in determining the strike and dip of the foliation. Individual strikes measured on the limb of a plication are commonly at right angles to the regional strike of the outer zone of gneissic and mylonitic rocks, although the regional strike within the zone of swirling follows closely the regional strike of the outer zone of foliated rocks (Fig. 6).

The other mappable feature is linear structure. In the case of the swirled gneisses this linear structure is determined by measuring the projected strike and pitch of the axes of the plications. The pitch is invariably in the same direction and at the same or a slightly lesser angle than the dip of the surrounding shell of gneissic and mylonitic rock. Linear structure determined by the axes of plications is most conspicuously developed where the outer zone of gneisses and mylonites turns sharply—here the regional foliation within the zone of swirled gneisses appears to have been squeezed into the axial portion of the warped outer zone of gneisses and to have shortened itself by the formation of innumerable plications whose axes plunge with the dip of the outer foliated bands (Fig. 6). Curiously, however, where the linear structure shown by swirl axes is well developed, the prevailing pitch of the axes is invariably northwest. This appears to indicate at least a slight regional control of this structure, for in areas where the dip of the outer zone of gneisses is to the west or southwest swirled structure with uniformly pitching axes is not found; here the swirling is either of ptygmatic nature, or is not developed.

Within the unswirled porphyritic gneiss of this facies a linear structure is also encountered in some areas. It is recognized by the orientation of hornblende needles, by the stringing out of feldspar phenocrysts, or by oriented clots of dark minerals, especially biotite. It corresponds exactly in orientation with the linear structure determined by plotting the pitch of swirl axes (Pl. 2).

Homophanous facies.—Traced eastward the plications and foliated structure of the swirled and porphyritic gneiss facies gradually diminish, and at two localities within the map area (Mount Hull and Tunk Mountain) disappear entirely so that the batholith rock becomes uniform

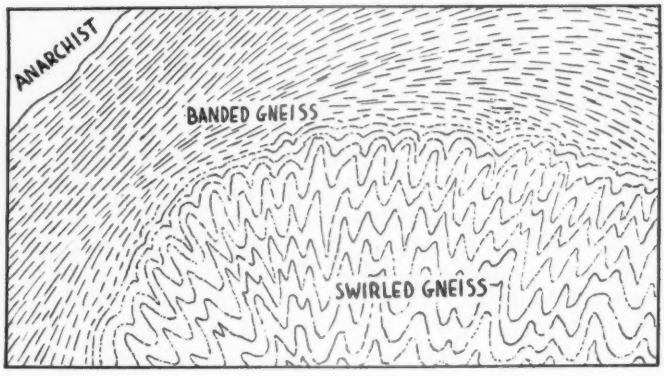


FIGURE 6.—*Schematic map of foliation*

At junction between Zone 1 and Zone 2 near Tunk Creek. Banded gneiss and swirled gneiss have identical regional strikes, though individual strikes of swirled gneiss vary widely.

petrographically and almost or quite structureless. Two other centers of structureless igneous rock are found in Mount Annie and Mount Bonaparte just off the map to the east, and reconnaissance investigations farther east indicate that much of the central portion of the batholith is made up of massive granitic rocks.

At both Mount Hull and Tunk Mountain the transition into this homophanous core from the swirled and porphyritic gneisses is very gradual. The rock gradually loses its foliation and other directional structures, evidences of cataclastic structure diminish and may even disappear completely until ultimately the rock gives way to massive unfoliated granodiorite. The material composing Tunk Mountain still retains traces of a linear structure shown by the alignment of hornblende needles, but the material of Mount Hull and Mount Annie appears to be entirely structureless. On the southwest side of Mount Bonaparte the transition from the porphyritic gneiss facies to a homophanous facies is very well shown. The porphyritic structure still persists in the homophanous rock, but the phenocrysts of microcline are euhedral and unbroken. However, to the west they gradually become cracked and strained, a faint foliation appears, and ultimately the rock grades into

porphyritic gneiss with augen-shaped microcline phenocrysts showing borders of crushed and albitized particles.

At some localities the foliation in the zone of swirled and porphyritic gneisses dips away from the cores of homophanous rock and the foliated portions of the batholith appear to surround the unfoliated portions as a girdle. Elsewhere, the transition to homophanous rock may take place directly along the strike of the foliation. Both on the western side of Mount Hull and the western side of Tunk Mountain the foliation actually dips inward toward the area of homophanous rock.

At a very few localities dikelike bodies of homophanous rock have transgressed the foliated border facies abruptly. Such relations may be seen on the southwestern side of Mount Annie, and less clearly on the wooded and poorly exposed western slope of Tunk Mountain. Irwin (1938) has described the truncation of an older foliated facies of the Colville batholith containing numerous strained phenocrysts by sharp dikelike masses of a younger finer-grained facies in the foundation of the Grand Coulee Dam about 40 miles to the southeast of this area. However, such cross-cutting contacts are exceptional; they were apparently developed because the homophanous facies completed its solidification a little later than the rocks of the bordering zones, and so are examples of autoinjection.

PETROGRAPHIC FACIES OF THE BATHOLITH

General statement.—So many kinds of rocks are found within the area mapped as Colville batholith that their systematic petrographic treatment is hardly possible without adopting a definite idea as to their origin.

Briefly stated, the events producing the diverse rock types of the Colville mass are believed to have occurred in the following sequence: (1) emplacement of a large granodiorite mass considerably below the levels now exposed by erosion; (2) extensive migmatization of the wall rocks (largely mafic igneous rocks, similar perhaps to the Anarchist greenstones) by this granodiorite; (3) almost complete solidification of a large part of the batholithic mass together with its migmatitic border; (4) rise of this almost solid mass to its present level in the crust accompanied by much granulation and mashing of the nearly solid border portion of the batholith and extensive brecciation of the wall rocks beyond the molar contact; (5) squeezing out, simultaneously with the rise of the mass, of the still liquid part of the border facies of the batholith to form the irregular bodies of pegmatite and alaskite; (6) partial to complete recrystallization of the crushed and granulated material; (7) final solidification of the central homophanous portion of the batholith and of the outfiltered pegmatites and alaskites. All of these processes overlap.

Individual rock names are based largely on texture. From the standpoint of the control of the texture three processes were vitally important: (1) solidification from a melt, (2) migmatization, (3) metamorphic changes brought about by cataclasis and recrystallization. Very few rock specimens from the Colville mass show the effects of only one of these processes to the exclusion of the other two, and the texture of many shows clearly the effects of all three. However, despite this overlap, it has been found convenient to group the materials from the Colville batholith into three petrographic facies. These petrographic facies are designated: (1) primary igneous facies, (2) migmatite facies, (3) autometamorphic facies. Each facies includes several kinds of material which differ in mineral composition, texture, or both.

In general the rocks belonging to the primary igneous facies make up that part of the batholith which for purposes of structural mapping has been designated the homophanous facies. In the form of alaskite and pegmatite they are also extensively developed in the gneissic and mylonite facies. They are also very important in the areas designated the swirled and porphyritic gneiss facies, but here the effects of cataclasis and recrystallization are also pronounced and in consequence this zone is largely composed of rock types intermediate between the primary igneous facies and the autometamorphic facies.

Rocks representing the migmatite facies are restricted very largely to the outer gneissic and mylonitic border, although they are also developed locally in the areas of swirled and porphyritic gneisses.

Rock varieties classified under the autometamorphic facies have their greatest development in the outer zone but are occasionally found in the zone of swirled and porphyritic gneisses.

Primary igneous facies.—The rocks which have been classed under the primary igneous facies show the common textures and minerals of normal igneous rocks. They comprise mainly granodiorite and quartz diorite with minor amounts of pegmatite, alaskite, and lamprophyre.

GRANODIORITE AND QUARTZ DIORITE: The areas designated as the homophanous facies of the Colville batholith in Plate 1 are made up of rocks belonging to the granodiorite and quartz diorite clans. Granodiorite greatly predominates.

Rocks from this zone collected on Mount Hull are mottled, light gray, and medium-grained. In some specimens parts of the rock have a slightly sugary texture reminiscent of aplite. Microscopically a typical specimen has a hypidiomorphic to allotriomorphic granular texture somewhat modified by a minor amount of cataclasis shown by wavy extinction and microshearing, and locally by granulation and minor rounding along

some of the crystal edges (Pl. 5, fig. 1). Plagioclase (An_{20}) is the most abundant mineral. It occurs in crystals commonly 1 mm. long but ranging up to 3 mm. Albite twinning is mostly absent, or else is found only in a part of the crystal. Some crystals show zoning with little range in composition. Quartz is nearly as abundant as plagioclase. It occurs mainly in large crystals (1 to 3 mm.) but is common as fine grains in the crushed material. Potash feldspar is less abundant. It occurs in large crystals and in tiny pockets and stringers. Some shows faint microcline twinning, but for the most part it appears to be orthoclase. Myrmekite is common along plagioclase-orthoclase boundaries, and is also developed in the fine crushed material. The plagioclase of the myrmekite has indices of refraction little different from the surrounding plagioclase. Biotite is the only mafic varietal mineral. It occurs in small ragged flakes. The crystals are usually straight, locally bent. Muscovite is also present in small amount, occurring both in feldspar crystals and as larger flakes between. Accessories include magnetite, large apatite crystals, grains of allanite, zircon in stubby prisms and as slender rods, a very little sphene. Alteration products are represented by sericite and chlorite, but the rock in general is remarkably fresh. A mode analysis is given in Table 2.

A specimen from Mount Annie (just east of the central part of the map area) of which a chemical analysis and mode analysis are given in Tables 1 and 2 is similar save for a marked increase of plagioclase with respect to potash feldspar. The rock from Mount Annie also shows little evidence of cataclasis and contains small yellowish garnets as an accessory.

Material from the homophanous zone on Tunk Mountain is of several different kinds. Much of the rock of this area is similar to that of Mount Hull and Mount Annie save for a diminution in the amount of quartz and potash feldspar and the appearance of hornblende as well as biotite. Near the summit and along the northwestern slope of the mountain hornblende increases, orthoclase almost completely disappears, and a great diminution of quartz occurs. This rock, of which chemical and mode analyses are given in Tables 1 and 2 is a quartz diorite.

Porphyritic granodiorites containing phenocrysts of microcline up to 2 inches in length occur on the southwestern flank of Mount Bonaparte and grade westward into the porphyritic gneisses of Zone II. The phenocrysts are euhedral but commonly show evidence of shearing strains by strain shadows and microfaults even when the groundmass of the rock is quite devoid of cataclastic structures.

FIGURE 1. HOMOPHANOUS GRANODIORITE
Mount Hull. Some cataclasis shown by streaks of granulated material, and by cracks and wavy extinctions. A few patches of flamboyant quartz. Crossed nicols, x 4.

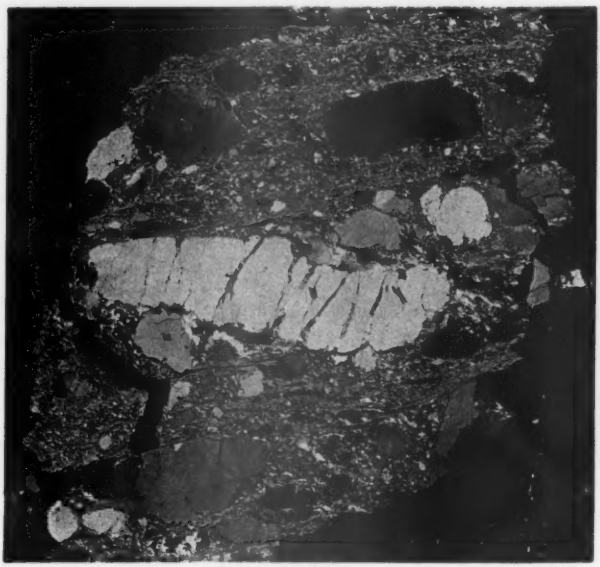


FIGURE 2. MYLONITE
Porphyroclasts in microbrecciated groundmass. Note quartz-feldspar mesostasis healing cracks in large porphyroclast. Crossed nicols, x 4.

PHOTOMICROGRAPHS OF ROCKS FROM COLVILLE BATHOLITH

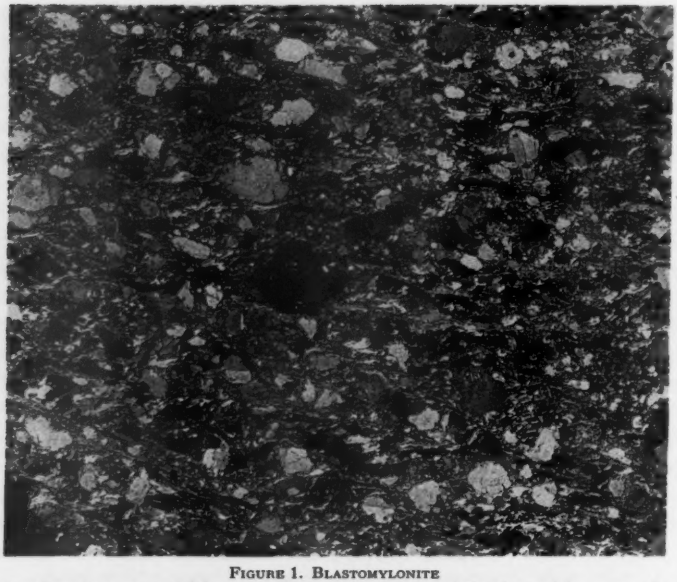


FIGURE 1. BLASTOMYLONITE
Porphyroclasts separated by intertwining shear zones partially obscured by recrystallization products and quartz-feldspar mesostasis. Crossed nicols, x 4.

FIGURE 2. BLASTOMYLONITE
Detail of a plagioclase porphyroclast from Figure 1. Crossed nicols, x 15.

PHOTOMICROGRAPHS OF ROCKS FROM COLVILLE BATHOLITH

Within Zone II (the swirled and porphyritic facies) granodiorites are also the prevailing rocks. Both hornblende-biotite granodiorites and biotite granodiorites are found, the latter in greater abundance. The chief petrographic difference between these rocks and those in the homophanous zone is the increasing evidence of cataclasis and recrystallization as one proceeds to the west. Cataclastic effects are usually visible in the field and are always visible microscopically.

Although in mineral composition rocks of this zone differ little from the granodiorites of the homophanous zone, they are invariably darker, more variable in texture, and show a foliation which in some cases is merely incipient, in others well developed. They are of various colors, but the most common rocks are greenish-black or dark gray intermingled with white forming mottled surfaces. They are commonly porphyritic, and the large phenocrysts of potash feldspar locally show more or less granulation along their edges. Individual phenocrysts may range in shape from perfectly euhedral crystals to lens-shaped augen surrounded by granulated fragments that form tails on either end. Under the microscope no essential difference in mineral composition from the rocks of the homophanous zone can be observed, but the fabric of the rock is different. One-third or more of the rock will be composed of fine-grained crystal aggregates which wander irregularly in swirls between the larger crystals, and yet are oriented into streaks which give the rock a definite foliation (Pl. 6). The larger grains, occurring either as individual crystals or in groups more or less isolated in this finer-grained matrix, show obvious evidence of rather intense cataclasis. They are cracked, bent and deformed (Pl. 4, fig. 3; Pl. 6, fig. 2). They show wavy extinctions, an abundance of microfaults, and their edges are commonly surrounded by angular fragments rubbed off by mechanical movement.

However, the streaky matrix in which these larger crystals are imbedded is not composed merely of a mylonitic paste formed by the mechanical milling down of the larger crystals. Although material of this derivation is present, the matrix in general is characterized by an almost complete absence of cataclastic structure. Broken up and granulated material, if once present, has been largely recrystallized into a quartz-feldspar-biotite aggregate. The areas suspected of being recrystallized mylonitic material show textures akin to metamorphic granulites. That is, the crystals are all of fairly uniform size, are entirely allotriomorphic, and occur in small grains whose boundaries may be sutured or fairly straight. Sieve texture may or may not appear. Imbedded within this granulitic matrix are large grains ranging in size up to the large crystals previously described, and showing crushing effects.

These inclusions, when composed of feldspar, are commonly partially converted to myrmekite, and grains and irregular areas of myrmekite are distributed all through the groundmass.

Whether the granulitic patches represent a mylonitic paste which because of the heat of this still partially unsolidified portion of the batholith recrystallized in the solid state, or whether they represent simply a part of the rock in which the individual fragments derived from cataclasis have been enlarged by accretion from the still fluid residuum is a problem that cannot be satisfactorily answered with the data at hand. Both processes were probably important, though recrystallization was very likely predominant.

Even more abundant in the groundmass than these areas which are believed to be due to recrystallization is a sort of mesostasis made up largely of feldspar and quartz, but containing varying amounts of biotite and other minerals. In this mesostasis potash feldspar is often as abundant and may exceed plagioclase in amount. The mesostasis wanders irregularly between the large cataclastically deformed crystals and enters and seals cracks and openings within them (Pl. 5, fig. 2). This material also appears to penetrate or sometimes to grade gradually into the fine-grained material formed by recrystallization. Unlike the recrystallized patches, however, this quartz feldspar aggregate does not have the typical granulitic texture. Biotite flakes within it are straight and euhedral; orthoclase, though without crystal form, has wholly irregular outlines with fingers and bulbous extensions quite unlike the discrete uniform-sized grains of the granulitic aggregates. Quartz may be present in shapeless grains, or more commonly is in long peculiarly shaped bladed crystals which may be intertwined with feldspar or may occur together in quartz aggregates. This material, because of its mobility as shown by its healing fractures in both the matrix and in isolated mineral grains, is believed to be the residuum of molten material which solidified in the matrix of the rock after crushing had ceased. The deformation responsible for the cataclastic effects in the larger crystals, therefore, must have occurred before complete solidification, and the crush structures are of protoclastic origin.

At all events, the rock described seems best classified as an igneous rock. Most of it has been formed by direct precipitation from a magmatic solution, but the course of the solidification was interrupted by a period of mechanical movement which deformed and partially granulated the early solidified components. These strained components were then partially recrystallized and were also cemented by final crystallization of the still liquid residuum. The typical swirled structures of this part of the batholith, described in an earlier section, confirm the

conclusion that the movement responsible for cataclasis took place in material not yet completely solid.

PEGMATITE AND ALASKITE: Orthodox pegmatite dikes are found occasionally in the interior of the batholith but account for little of its volume. Along the border, however, are found huge masses of a peculiar rock with the composition of pegmatite, but with a texture more nearly that of a coarse normal granite. In these rocks plagioclase and potash feldspar are usually present in about equal proportions, though this ratio varies somewhat. They are therefore quartz monzonite pegmatites in composition. Some of the finer-grained material might more appropriately be called alaskite, but most of it is sufficiently coarse-grained to make the name pegmatite more suitable.

Pegmatite is almost invariably found along the border in small amount, but locally large volumes occur. The largest single mass underlies the metagabbro of Earl Mountain and is apparently continuous with smaller outcrops isolated by glacial debris west of Okanogan River. Enormous volumes of pegmatite are also found all along the north and northeast borders of the Mount Hull portion of the batholith where they may form an advance outpost in the direction of movement of the batholith.

Although most of the pegmatite is found at the contact, considerable volumes also occur as sill-like bodies and lens-shaped masses enclosed between the foliation planes of the gneissic and mylonitic facies of the batholith, and as irregular dikes cutting across the foliation. Locally, as near the mouth of Tunk Creek, great podlike and irregular masses of pegmatite are massed near the junction between Zone I and Zone II of the batholith. Sinuous dikelike bodies cut the porphyritic gneisses of the lower part of Bonaparte Creek.

The similarity in composition between the pegmatite and the feldspar-quartz mesostasis described as forming a cement for the cataclastically deformed crystals of the granodiorites of the swirled and porphyritic gneiss facies is obvious, and the two are believed to have essentially the same origin. The more pronounced protoclastic movement occurring near the walls of the batholith, however, has in large degree expelled the residual uncrystallized mesostasis completely, and it has collected into irregular pegmatitic pools along the border of the batholith. Evidence of similar filtration differentiation on a smaller scale is frequently encountered in the swirled and porphyritic gneiss facies where small pools of the quartz-feldspar mesostasis have aggregated as irregular masses within the rock, or have even formed veinlike bodies with welded contacts which lie within the foliation planes or else wander irregularly across them. These relations are particularly common where the gneisses are intricately swirled. Pegmatitic material may collect along the axial

planes of the sharper swirls until nearly all vestiges of foliation disappear (Fig. 7).

Additional evidence of this mode of origin is seen in the petrographic character of the pegmatitic material. Under the microscope the pegma-

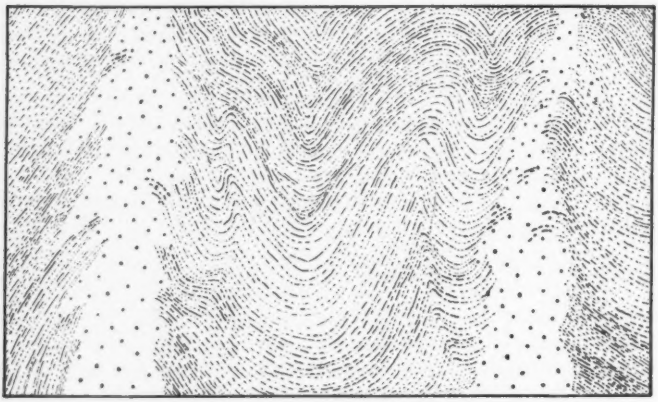


FIGURE 7.—*Pegmatite along axes of folds*
In swirled gneiss. Chewilikan Valley. Three-fourths natural size.

tites, though variable, are light-colored, coarsely granitic rocks, similar in many respects to the granodiorites of the homophanous facies of the batholith, but containing less plagioclase and biotite and much more potash feldspar. The plagioclase may have the same range in composition as that of the granodiorite or may be more sodic. The potash feldspar is more commonly microcline or microperthite. While the granodiorites are in general notable for the freshness of their minerals, feldspars and mafics of the pegmatites are commonly decomposed. In hand specimens the crystals are outlined only vaguely, and often the former presence of mafic minerals is indicated by gray-green patches which are resolved by the microscope into scattered wisps of chlorite. At a few localities alteration of the rock has proceeded so far that it shows no crystals megascopically, is a uniform light gray green, and resembles a light-colored chert. On the other hand, some pegmatite bodies are almost unaltered; these are more likely to be found in small dikes or minor injections within the batholith and not in the great pools at the molar contact.

Next to feldspar, quartz is the most abundant constituent of the pegmatite, and commonly accounts for 40 per cent or more of the total

mineral composition. Quartz occurs as large somewhat strained crystals and even more abundantly as the long-bladed crystals in swirled aggregates which are typical of the mesostasis in the partially crushed granodiorites.

Myrmekite is almost invariably present, usually showing rounded outlines toward bodies of potash feldspar that act as host. Biotite is present as ragged flakes, which in the larger pegmatite pools are usually altered to chlorite. Muscovite is commonly present in small amount but in some masses is lacking. The accessories are the same as those of the granodiorite: zircon in curious pencil-like crystals, apatite, magnetite, garnet, allanite, sphene, and locally epidote and tourmaline. In general the proportion of accessories is somewhat less than in the granodiorites; this is especially true of allanite and sphene, but the small yellowish to pinkish garnets as well as epidote and tourmaline are more abundant than in the granodiorite.

Small pegmatite dikes entirely free of crushing effects are common, but many areas of this rock show a minor amount of cataclasis. Locally, however, pegmatite dikes have been smeared to ribbons by mechanical movement. Cataclasis is more commonly indicated by abraded crystal edges, microfaulting, wavy extinction, and occasionally by accumulations of finely granulated material mixed with the products of recrystallization or the final residues of solidification.

Some of the pegmatite bodies contain numerous inclusions, consisting of highly crushed fragments from the gneissic and mylonitic zone. Fragments of the dark-colored migmatitic rocks of this zone are common, and at some localities they appear to be undergoing a marked secondary shattering and disruption within the pegmatite, accompanied by softening and metasomatic change. Fragments from the adjacent wall rock are locally common.

LAMPROPHYRE: Lamprophyre dikes associated with the Colville batholith are extremely scarce. With two minor exceptions, they are restricted in this part of the batholith to a small area near the summit of Mount Hull. The numerous lamprophyres cutting the hybrid syenite in the valley of Tonasket Creek are more probably associated with the Osoyoos batholith than with the Colville batholith.

The swarm of small (up to 5 feet thick) dikes on the upper slopes of Mount Hull consists of dark, fine- to medium-grained rock with considerable diversity of mineral content and texture. The most common variety is composed chiefly of markedly zoned plagioclase and light brownish-green hornblende, with minor amounts of brown biotite. The rock in the larger dikes is massive, that in the smaller ones often some-

what sheared. The principal trend of the dikes parallels a major joint system of the batholith—N. 60° E., with steep dips to the northwest.

Migmatite facies.—Much of the rock now found within the zone designated the gneissic and mylonitic facies is believed by the writers to have once been a migmatite developed through injection and reaction between the Colville granodiorite and wall rocks composed largely of mafic igneous rocks or their metamorphic derivatives. However, this border zone was so profoundly affected by cataclasis and recrystallization during the rise of the batholith that actual evidence of migmatization has been very much obscured, and the relations are difficult to interpret. Clear cut examples of the formation of migmatites are to be seen at only a few localities; for the rest of the zone conclusions must rest upon less direct evidence.

Greenstone-Granodiorite Migmatite: Indubitable evidence of migmatization of greenstone by the Colville granodiorite may be seen in exposures along the walls of Siwash Creek east of Antoine Valley, and on the divide between Siwash and Bonaparte creeks. Here a roof pendant of greenstone has survived erosion. Some distance from the contact this greenstone is rather massive with only a faintly developed schistosity. Closely spaced hackly joints break it into small blocks which accumulate as rubble over the steep slopes. As the contact is approached, however, the character of the greenstone changes markedly. The hackly jointing largely disappears, and an indefinite streakiness parallel with the contact appears and increases in intensity. Small porphyroblasts of potash feldspar appear in abundance and increase in size and amount nearer the batholith. The matrix of the rock coarsens, evidence of relict volcanic structures disappears, and ultimately by this increase in grain, coupled with the increasing streakiness, the rock begins to take on the appearance of a coarse-grained melanocratic gneiss. Cataclastic structures are also increasingly common, and locally mask the previously described coarsening. Veins and stringers of the granodiorite penetrate into even the hackly jointed, mildly metamorphosed greenstone, but in the streaky coarse grained porphyroblast-rich types they become very abundant. The granodiorite has penetrated as sill-like bodies parallel to the streaking of the country rock. Contacts between it and the coarsened greenstone are commonly indefinite but may locally be sharp. Usually there is a marked coarsening of the greenstone and a great increase in the number of potash feldspar porphyroblasts adjacent to the granodiorite sills. With increase of granodioritic material the slablike bodies of migmatized greenstone begin to break up into fragments which are pulled out, or partake of the swirling movements

that have affected the granodiorite. At the same time they coarsen in texture, and become medium- to coarse-grained amphibole-rich rocks in which all vestiges of the original volcanic texture have disappeared. Extensive cataclasis in some places reduced them to still finer materials.

At most localities, however, the outcrops do not show such a complete transition between greenstone and granitic rocks. On much of the ridge between Siwash and Bonaparte creeks, and in the lower portion of Bonaparte Valley granodiorite occurs intimately mingled with rocks corresponding in composition to hornblende schist or to dioritic rocks rich in hornblende. These amphibole-rich materials occur as long slab-like inclusions or as wavy stretched bands between the layers of granodiorite, and where cataclasis has not been profound these zones of mixed rock are exactly similar to the more thoroughly migmatized portions of the area on Siwash Creek. Commonly, however, marked cataclasis has affected nearly all portions of this zone and has greatly obscured the relations. Materials probably once representing greenstone-granodiorite migmatites now form banded recrystallized rocks in which granulitic materials composed largely of hornblende with minor amounts of feldspar and other minerals alternate with or grade into bands composed of granodiorite and of mylonite or its recrystallized equivalents. Without the evidence of gradual transformation of this material to the migmatites of Siwash Creek, the hypothesis that these cataclastically deformed and recrystallized materials were derived from granodiorite-greenstone migmatites would have to rest solely on their heterogeneity of composition, and their peripheral distribution.

Most of the material thought to be of migmatitic origin is believed to result from the migmatization of greenstone or other mafic rocks by granodiorite. Throughout most of the zone the proportions of granodiorite exceed those of the greenstone, but these relations may be reversed as the molar contact is approached, and locally elsewhere. Thus in the bed of Tunk Creek about 4 miles above its junction with the Okanogan, an outcrop of amphibolite schist occurs containing minor amounts of garnet, pyroxene, and feldspar, injected by ramifying quartz-orthoclase veinlets, but otherwise containing very little granodioritic material. This probably represents a roof pendant of greenstone thoroughly permeated and recrystallized by the granodiorite, but not intimately shattered and interpenetrated by it. It has been mapped as a portion of the batholith, however, because of the fact that the rock has been sufficiently softened to partake in some degree of even the intimate swirling that characterizes this portion of the adjacent igneous mass.

Locally rock varieties are found which may represent the migmatization of materials other than mafic igneous rocks. Along lower Tunk

Creek, and on the adjacent ridge which separates Wagonroad Coulee from the Okanogan River occur gneissic rocks very rich in biotite, possibly representing the products formed as a result of the migmatization of phyllites or other pelitic rocks by the batholith. With these biotite-rich gneisses occur blocklike and tabular inclusions of rock composed mainly of quartz, but containing minor amounts of feldspar and myrmekite, which may once have been bands of quartzite in the original phyllitic sequence. Even more puzzling is the occurrence of thulite-bearing gneisses along Tunk Creek. The thulite is intergrown with quartz, and occurs in bands and smooth podlike masses enclosed in the foliation planes of highly cataclastic biotite gneisses.

Locally, within the mapped area the gneissic and mylonitic zone is over 2 miles thick. This great thickness might suggest that the rocks of this zone have not developed from migmatites. However, the adjacent Chelan batholith, believed to be of approximately the same age as the Colville, is intruded into much the same kind of wall rocks and is bordered by an astonishingly thick zone of contact breccia and migmatite—a zone that locally is over 4 miles thick (Waters, 1938). This border zone of the Chelan batholith is believed to represent almost exactly the condition of the margin of the Colville mass just prior to the forcible uptrusion that formed the structures due to cataclasis and recrystallization.

Furthermore, in the Kettle Range the development of a very extensive zone of migmatitic rocks in the Colville batholith has been described by Campbell (1936). At this locality the late rise of the mass, though producing many of the same structural features observed in the Okanogan Valley, has not so completely obliterated early textures and structures as it has along that portion of the mass described in this report.

In summary then, the progressive stages in the migmatization of greenstones by granodiorite are well displayed at one locality, and can be closely inferred at many others. Portions of the batholith outside the mapped area as well as the adjacent contemporaneous Chelan batholith show widespread migmatization. In much of the area under consideration, however, the original migmatitic character of the rock must be inferred by its similarity in composition to migmatite, by its streaky development of bands which vary in composition from greenstone to granodiorite, and by the transition of this kind of material into rocks which are recognizable migmatites. In a few places migmatization of pelitic sediments is inferred.

Osoyoos-Colville-Greenstone Hybrid Syenite: One of the most complex areas mapped for this report was encountered where the Osoyoos and Colville batholiths join. The zone between the batholiths at this locality contains very abundant pegmatite, some of it deficient

in quartz. The migmatite portion, that is, the gneissic and mylonitic facies of the Colville batholith, thins toward this locality and is generally absent. Just beyond what is interpreted as the molar contact of the

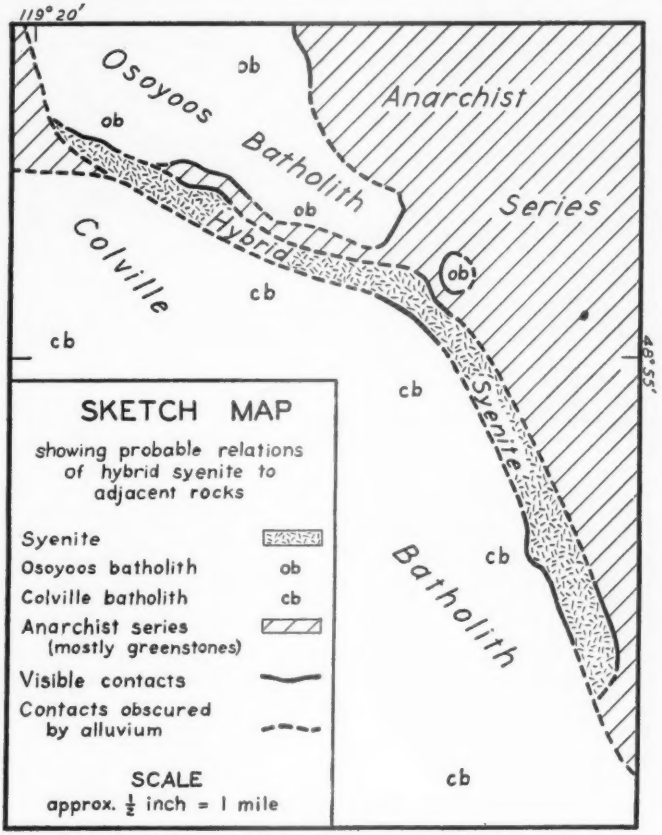


FIGURE 8.—*Sketch map of hybrid syenite*

batholith appears a mass of heterogeneous dominantly syenitic rocks which border the Colville batholith along Tonasket Creek and a part of Eden Valley. This mass consists in general of peculiar rock types which cannot be definitely assigned to either batholith. For this reason, and also because the syenite is largely covered by alluvium and glacial debris, it has not been shown on Plate 1, but Figure 8 has been prepared

to indicate visible and inferred contacts of the syenite with adjacent rocks. In general the syenites are gneissic hornblendic rocks ranging in composition from mafic to felsic types. They are saturated with pegmatites both in the form of definite cross-cutting dikes and as bodies

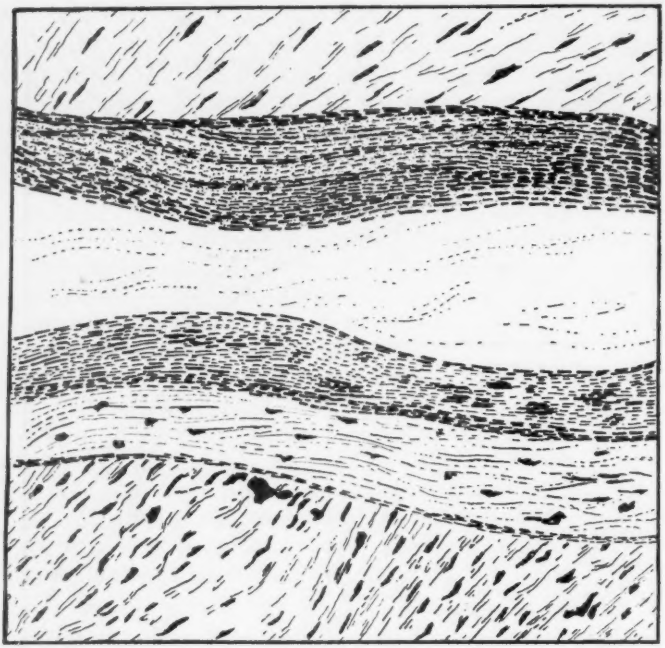


FIGURE 9.—*Intensely sheared syenite*
With sheared pegmatite dike in center. Syenite with swirled foliation at edges. Natural size.

into which the syenite appears to grade. The structure of the syenite is unbelievably complicated by swirls, faults, and zones of shear (Figs. 9, 10; Pl. 8).

The mafic varieties of hornblende syenite are best displayed in a road cut about 1 mile north of Tonasket Creek and 7 miles east of Oroville. Foliation in these rocks is fairly even, curving slightly near some dikes. Pegmatite, mostly a variety low in quartz, cuts the rock in all directions as dikes ranging from a foot or more wide down to the tiniest stringers. In places (Pl. 8, fig. 1) the dike material seems to be forcing

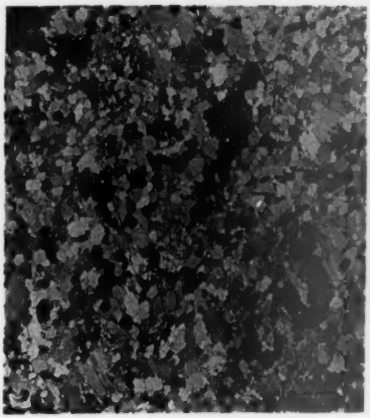


FIGURE 1. GRANULITE
Recrystallized mylonitized granodiorite from periphery of Zone 2. Crossed nicols, x 4.

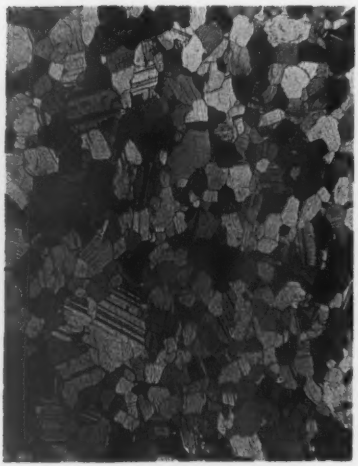


FIGURE 2. GRANULITE
Recrystallized migmatitic rock from the gneissic and mylonitic zone. Crossed nicols, x 8.

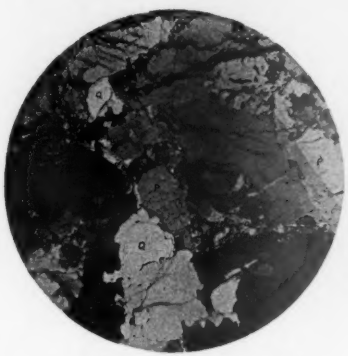


FIGURE 3. HOMOPHANOUS GRANODIORITE
K is potash feldspar, P is plagioclase, Q is quartz. Note myrmekite embaying potash feldspar. Crossed nicols, x 24.

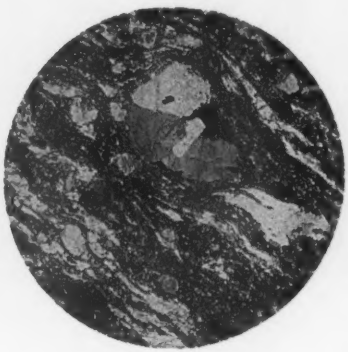


FIGURE 4. MYLONITE
Extreme crushing, minor recrystallization. Note feldspar porphyroclasts and secondary flamboyant quartz. Crossed nicols, x 8.

PHOTOMICROGRAPHS OF ROCKS FROM COLVILLE BATHOLITH

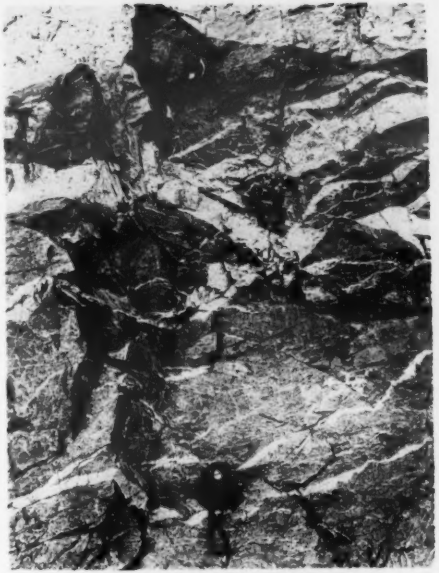


FIGURE 1. PEGMATITE AND SYENITE
Angular fragments of mafic syenite separated by quartz-poor pegmatite. Small dikes at bottom of picture show slight dislocations healed by later movements of syenite.

FIGURE 2. HYBRID SYENITE
Showing plexus of small and large shears, many partially or completely healed. Note that pegmatite dikelet in center is younger than some shears but is cut by others. Specimen 16 inches long.

ROCKS FROM COLVILLE-OSOYOOS BORDER ZONE

the syenite apart in great angular chunks. Later movement has faulted some of these dikes, often in steplike fashion, and locally the surrounding syenite was still plastic enough to well in and fill the fractures in the dikes (shown by the two lower dikes in Figure 1 of Plate 8).

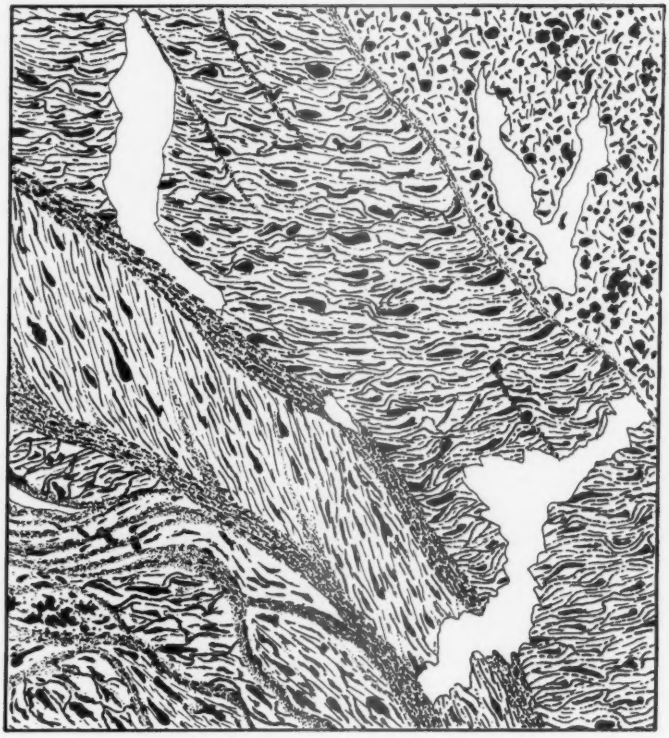


FIGURE 10.—*Complicated shears and swirls in syenite*
One-fourth natural size.

More complex structures are shown by the siliceous gneissic syenite at several points. The gneissic foliation of these rocks is swirled to an extreme degree; its pattern in places resembles eddies in a turbulent stream. Through it wind stringers of pegmatite, and lines of fine, dark material which represent inter-twining minor zones of shear. Something of this complexity is suggested in Figure 10 and Figure 2, of Plate 8, but no photograph or drawing adequately portrays the myriads of con-

torted lines. In addition to the ubiquitous pegmatite stringers the rock is cut by larger pegmatite dikes, mostly of a variety with conspicuous blue-gray quartz and tiny red garnets. Some dikes are quite undeformed, others are sheared into fine augen gneisses containing prodigiously elongated minerals; the adjacent gneissic syenite is in places sheared to gray-green mylonite. Shear zones of several ages are represented—some healed by plastic flow of the surrounding syenite, others sealed by granulitic textured rocks formed by recrystallization, younger ones lined with mylonite showing little trace of recrystallization, and still younger ones with prominent slickensides.

Where contacts of these syenites with either the Colville or Osoyoos batholiths or the metamorphic rocks of the Anarchist series are exposed, or nearly exposed, the actual line of separation is almost always obscured by masses of pegmatite, lamprophyre dikes, or both. About a mile west of the previously described road cut the mafic syenite at the tip of a ridge appears to grade northward into normal Osoyoos rocks, but the abundant pegmatites and poor exposures make the transition not absolutely certain. A syenite-Anarchist contact exposed in the northwest part of sec. 36, T. 40 N., R. 28 E., also appears to be gradational. A traverse across this zone shows phyllites with increasing grain size; the biotite giving place to hornblende, diopside, and epidote; feldspar increasing in size and amount; until at length appear intercalated syenitic bornblende gneisses of various types in narrow bands which become gradually less finely foliated and more uniform in composition.

Syenite-Colville contacts are uniformly obscured by pegmatite or covered by alluvium, but at several localities the two rocks can be located very close to one another. At some points the rocks appear to be entirely distinct, at others the relations are most confusing but cannot strictly be called gradational.

Pegmatite commonly crosses the syenite in definite dikes; elsewhere its relations to the syenite are gradational, and at still other localities (Pl. 8, fig. 1) the syenite has welled into fractures in pegmatite dikes. The syenitic character of some pegmatites at this locality has already been commented upon. Summing up the conflicting evidence it appears that there is no definite evidence of wholesale mixing of syenitic and Colville rocks on the one hand, nor evidence of any considerable difference in fluidity between them on the other.

As far as their distribution is concerned the syenites appear to be more closely related to the Colville than the Osoyoos batholith (Figure 8). The syenite zone throughout its entire extent (a distance of 8 miles) lies along the edge of the Colville batholith, and follows the curve in the batholithic contact to the northeast of Mount Hull. It is perhaps

significant that where the syenite rims the batholith the gneissic and mylonitic rocks of migmatitic derivation, which elsewhere in this area form a continuous shell about the Colville mass, thin and disappear completely, bringing the more homogeneous rocks of the homophanous facies and the swirled and porphyritic gneiss facies directly against the syenite.

On the other hand the syenite is separated from the Osoyoos batholith at most places by a thin tongue of Anarchist metamorphic rocks, and in the area east of Mount Hull the syenite continues southeastward some 4 miles beyond the last exposure of Osoyoos rocks. However, at the other end of the syenite the gradational contact with the Osoyoos was found.

Petrographically rocks of this zone are gneissic hornblende syenites varying considerably in their content of dark minerals and showing other inhomogeneities. They have little apparent relation to the staple rock varieties of either the Osoyoos or Colville batholiths. Potash feldspar is much more abundant than in either batholith, quartz much less abundant. Plagioclase is a little more sodic than in Colville rocks, considerably more sodic than in Osoyoos rocks. Hornblende is the only mafic mineral; in the siliceous varieties it occurs in peculiar but characteristic broad spindle- and tadpole-like shapes. Epidote is uniformly present, but abundant only in occasional large clots and veinlets. Accessories (sphene, magnetite, apatite, zircon) are abundant, and include garnet in the mafic varieties. Gneissic structure is well developed and fairly even in the hornblende-rich varieties, less clear and often highly swirled in the feldspathic types. Textures may be hypidiomorphic granular, but more often range from one suggesting the mosaic textures of metamorphic rocks to a strongly cataclastic texture approaching that of a mylonite.

Origin of the Syenite: From the conflicting and confusing evidence presented above four hypotheses as to the origin and relations of the hornblende syenite are open: (1) The syenite is a separate intrusion not directly related to either batholith; (2) it is a marginal facies of the Colville batholith; (3) it is a marginal facies of the Osoyoos batholith; (4) it is a hybrid rock formed by the action of emanations and magmas from both the Osoyoos and Colville batholiths upon a thin septum of Anarchist rocks, or upon the migmatitic rocks which elsewhere form the outermost zone of the Colville batholith. The first hypothesis is improbable because of the peculiar form of the intrusion, because of contact metamorphism far more intense than would be expected from such a small mass, because of its intimate structural relations to both batholiths, and because of locally gradational relations to them. An

altogether satisfactory decision between the remaining three hypotheses does not appear to be possible. For the following reasons the fourth hypothesis of hybridism is favored by the writers as the most logical explanation of the complex relations.

Chemical analyses are not available, but considerable information about the chemical composition of the syenite is brought out by the petrographic study. Evidently the chief chemical difference between the granodiorites of the Colville and Osoyoos batholiths as compared to the hornblende syenites is lower SiO_2 and higher Na_2O in the latter. Probably K_2O is also higher in the syenites, although the absence of biotite must in part compensate for the much greater abundance of potash feldspar. No recognized process of crystal differentiation will bring about an increase of silica and simultaneous decrease of soda and potash, so an explanation involving assimilation seems more reasonable. Immediately at hand for assimilation are great quantities of spilitic greenstones which would have precisely the desired effect on a granodioritic magma or the pegmatites derived from it. Furthermore, along the strip of Anarchist rocks separating the syenite from the Osoyoos batholith amphibolites and fine-grained gneissic syenites are intimately interlayered.

One difficulty presents itself: Why should hornblende syenites develop only along this portion of the granodiorite contact and not at several other localities where these two varieties of rock meet? Evidence presented before suggests that the Colville and Osoyoos batholiths were plastic almost or quite simultaneously, and this evidence is further elaborated in another paper (Krauskopf, 1941). The obvious answer to the local development of the syenites is that here, where the two batholiths collided, there was a greater concentration of heat and volatiles, a supposition supported by the abundance of pegmatite and lamprophyre but not capable of direct proof.

Significantly, at Earl Mountain, the only other locality along the batholithic contact where pegmatites are particularly abundant, definite hybridism of gabbro by pegmatite has taken place. Here, because of the lower temperature, hybridism did not advance to the stage of converting gabbro into syenite; the process was essentially a high temperature metasomatism resulting in introduction into the gabbro of biotite and especially potash feldspar which varies in amount in different specimens and becomes more abundant as the pegmatite is approached. The potash feldspar is present as tiny pockets, stringers, and veinlets cutting the other minerals. Accessories are also present in this metagabbro in greater amount than normal, a condition that particularly characterizes the hornblende syenites.

Additional evidence of the generation of syenitic rocks as a result of the hybridism of melanocratic igneous rocks by pegmatite can be seen in the deep coulee notches just southeast of the mouth of Tunk Creek, a locality many miles from the two areas just discussed. At this point great volumes of pegmatite have welled into the zone which separates the highly migmatitic gneissic and mylonitic facies of the Colville batholith from the swirled and porphyritic gneiss facies. The mafic rocks in the gneissic zone have been split up and locally shredded by the enclosing pegmatite, and are in all stages of dissolution. Those bodies of pegmatite particularly rich in partially dissolved xenoliths are universally lower in quartz, and many approach the hornblende syenites in composition, but lack the structural complexities of the hornblende syenite zone. The attack upon the mafic migmatitic border of the batholith by its own pegmatitic juices resulting in the formation of syenitic rocks on a small scale at this locality is clearly demonstrated. The complete absence of this migmatitic border in the locality where the hornblende syenites are developed, together with the presence of the great volumes of pegmatite, and the occurrence of the syenites as a narrow strip between two great batholiths which were emplaced approximately simultaneously, is thus believed to favor the hypothesis of their origin by hybridism. As a result of the collision, or near approach of the batholiths, the liberation of considerable heat might take place, leading to the impregnation of the greenstones between the batholiths with active solutions and pegmatitic magmas and their conversion to semifluid syenite. The newly formed syenite and accompanying pegmatitic effusions escaped, as the batholiths were crowded together, along the margin of the Colville mass. Subsequently, as cooling proceeded, the syenite belt was squeezed and contorted by further movements in the batholiths, giving it the singularly complex structural relations observed today. At different times during this process dikes were intruded into it, the early ones suffering from continued shear, the latter ones remaining undeformed.

Autometamorphic facies.—In earlier sections of this report it has been mentioned that much of the gneissic and mylonite facies is made up of rocks with cataclastic and crystalloblastic textures. Indeed, from a petrographic point of view, much of this portion of the batholith conforms more closely to a mass of metamorphic rocks than it does to an igneous body. However, the rocks with metamorphic textures are so closely related to rocks of definite igneous aspect, and have so clearly had their metamorphic textures imposed upon them by processes associated with the rise of the batholith, that there is no justification for

assuming that the metamorphic features were produced long after batholithic solidification. The metamorphic textures are due to autometamorphism brought about by protoclasis followed by local recrystallization as the almost solid igneous body rose and pulverized its own nearly crystallized border.

During the rise several definite autometamorphic rock types were produced. Immediately at the molar contact mylonites with porphyroclasts of potash feldspar are abundant. Local areas of protomylonite and some rare thin bands of lustrous material approaching pseduotachylyte, are present. These rocks are absent or poorly developed in that part of the batholith margined by the hybrid hornblende syenites, and in general are less conspicuously developed in the area north of Mosquito Creek than they are to the south. Magnificent exposures in three dimensions may be found on the east bank of the Okanogan River just downstream from Riverside and in McLaughlin Canyon and the area west of it. The mylonites usually show effects due to recrystallization in greater or less degree, and commonly grade into rocks best described as blastomylonites. These in turn grade into coarser-grained rocks with granoblastic texture which may or may not show a gneissic banding, and which are accordingly called granulites and granulitic gneisses. All of the various metamorphic rock varieties mentioned, and particularly the more siliceous granoblastic types, show every transition into definite igneous rocks, and indeed are in many cases so permeated with magmatic material as to make their classification as igneous or metamorphic a purely arbitrary matter.

MYLONITE[3] AND PROTOMYLONITE: Near the molar contact are light-colored to almost black rocks with a streaky irregular gneissic foliation whose bands may be clearly continuous and parallel, may be lenslike and wavy, or may crisscross each other at highly acute angles in the most intricate fashion. Set in the matrix of the rock are potash feldspar augen, or more rarely lens-shaped porphyroclasts of other minerals. The feldspar augen are commonly from one-fourth inch to an inch long and are often surrounded by a film of broken particles rubbed off by friction and drawn out into trails at either end of the eye. Locally augen are inconspicuous or absent.

These rocks resemble ordinary augen gneisses save that the matrix is in many cases so fine-grained that its constituent materials cannot be identified even with a hand lens. Microscopically, this matrix instead of showing the typical crystalloblastic texture of dynamic metamorphic augen gneisses is characterized by the cataclastic structures typical of

[3] For a glossary of the terms that have been adopted in describing the cataclastic rocks of this region see Waters and Campbell (1935, p. 477-481).

a mylonite. The banding so conspicuous in hand specimen is due to variable degrees of crushing in different layers. Some thin layers are so intimately crushed that even under magnifications of 100 diameters the grains are scarcely visible, whereas adjacent bands may show fragments averaging one fifth of a millimeter or more in diameter. There is also considerable variation in size of particles within individual bands. As in most areas where mylonites are developed every stage in the process of crushing from the development of protomylonites with incipient mortar structure to the complete streaking out of exceedingly finely crushed material may be seen. The separate steps in this process have been well described by many authors (Quensel, 1916; Backlund, 1924; Waters and Campbell, 1935) and need not be repeated here. Various stages are illustrated in Plates 5, 6, and 7.

However, the mylonitic rocks from the Okanogan Valley show certain structural and petrographic features not commonly seen in mylonites from other parts of the world. They are not related to thrusts, but were developed along the periphery of the batholith as it moved upward against a stationary wall. Furthermore, evidence clearly shows that at the time of their development magmatic juices and fluids still formed at least a part of the rock that was being deformed; rock flowage was accomplished in a crystal mesh which still contained a small proportion of interstitial liquids. These liquids have been squeezed out into bodies of pegmatite which have all shapes and dimensions. Some of them cross the foliation planes produced by shear banding, others are concordant with them, and still others lie at various angles or have very irregular forms. Many of the pegmatites cutting the foliation have in turn been deformed by a continuation of foliation plane slip. Moreover, the simultaneous expulsion of the pegmatites and development of the crush structures in the border zone of the batholith is shown by the fact that some pegmatites have been sheared to ribbons and drawn out into banded mylonites; these are crossed by younger faulted pegmatites showing mortar structure. They in turn may be cut by younger undeformed pegmatite bodies, which may nevertheless show irregularites in trend (Fig. 3) produced by foliation plane slip in the adjacent mylonitic gneiss before the pegmatite had solidified.

Although the complex relations of pegmatites to mylonitic rocks can best be seen in large outcrops the same phenomenon is also commonly displayed even in individual thin sections cut from rocks of the mylonitic zone. Microscopic study reveals a quartz-orthoclase mesostasis permeating the rock, filling cracks and openings in broken porphyroclasts, and wandering along the crush bands. Locally this mesostasis collects in pools from which irregular ramifications go forth into the pulverized

rock. The quartz-orthoclase liquid seems to have been active chemically and apparently has greatly aided recrystallization of the adjacent mylonitic paste. In fact the veinlets of quartz and orthoclase are commonly much better developed in the blastomylonites than in the true mylonites. Widespread development of myrmekite in the crush zones or along the borders of potash feldspar porphyroclasts, and local albitization of granulated borders of orthoclase porphyroclasts, are additional evidences of solutions and magmatic juices in the crush rocks.

Although decidedly variable, the large bodies of pegmatite seen in outcrops commonly differ slightly in composition and texture from the quartz-feldspar mesostasis permeating the crush rock. The pegmatites nearly always contain considerable though variable amounts of plagioclase and a small proportion of ferromagnesian minerals. The mesostasis usually lacks these minerals, though some albite may be present. Furthermore, the quartz of the mesostasis commonly occurs in long, relatively straight crystals intricately intertwined with each other yet preserving a fairly definite parallelism. This character suggests the structure of "comb quartz" from veins save that the direction of crystal elongation is more commonly parallel with the length of the "vein" than across it. These elaborately intertwined crystals are commonly quite free of orthoclase but may be intergrown with it. As the proportions of orthoclase in the mesostasis increase the quartz assumes a more normal habit. These features suggest that the aggregates of elaborately intertwined quartz crystals may have been deposited from solution, and that the mesostasis represents a slightly more advanced stage in differentiation than the larger bodies of liquid expelled as pegmatites.

Blastomylonite and Granulitic Gneisses: Only rarely are mylonitic rocks found completely free of new minerals developed by recrystallization of the cataclastic paste. In most of the crush rocks minute scales of biotite have developed in considerable amount, and in properly cut sections they may give the rock a marked regional pleochroism. Other secondary minerals commonly developed are quartz, feldspar (largely albite and orthoclase), hornblende, and muscovite. Chlorite, epidote, and sphene are locally developed, and may be prominent in individual specimens but are not widespread.

In some specimens these new minerals developed by recrystallization are only sparingly present and make up perhaps less than 10 per cent of the mylonitic rock. Much more common, however, are rocks whose matrix consists largely of products of recrystallization and which are best described as mylonite gneisses rather than mylonites. These in turn grade into blastomylonites with a definitely crystalloblastic matrix,

their derivation from mylonitic rocks is shown only by the presence of scattered porphyroclasts or other inherited fragments that still preserve cataclastic effects.

Petrographically the blastomylonites are usually medium-gray rocks with a distinct sugary texture in hand specimen. Microscopically the mineral grains are seen to interlock in typical crystalloblastic fashion; in some cases the boundaries are sutured, in others they form a uniform mosaic somewhat similar to the texture of aplites. Sieve textures are also common. Hornblende and biotite are often aligned and may give the rock a rude foliation; in other cases foliation is due to the alternation of crystalloblastic bands with layers of quartz-feldspar mesostasis, with layers of incompletely recrystallized mylonitic material, or with intricately intermingled areas of both.

Cataclastic effects, at least in minor amount, are nearly always present in the blastomylonites. Furthermore, in thin sections evidence can sometimes be seen that the processes of cataclasis and recrystallization overlapped somewhat. New minerals have been broken and pulverized by cataclasis, and sometimes in turn recemented by infiltration of the liquid quartz-feldspar mesostasis. However, in general, recrystallization continued after cataclasis had ceased and has resulted in complete obliteration of the original cataclastic structure. The newly formed quartz may even be totally devoid of strain shadows.

With increasing distance from the contact recrystallization effects become more and more pronounced and the blastomylonites achieve a coarser grain. Eventually all or nearly all traces of cataclasis disappear and the rock coarsens until it approaches the grain size of the normal granodiorite. These granoblastic rocks may or may not show a foliation. The unfoliated ones are identical with normal granulites (Pl. 7, fig. 1), the foliated types may be described as granulite gneisses.

Many of these granulites and granulite gneisses present textures which are neither definitely igneous nor definitely metamorphic. Others are clearly crystalloblastic with sutured boundaries between the minerals, and definite sieve textures with every mineral enclosing every other mineral. Locally, however, the cores of feldspar and hornblende crystals will show sieve effects, but the outer periphery of each mineral is made of a relatively inclusion-free band which appears to have been deposited by a magmatic solution. In a very few cases quartz crystals show strain shadows in the core but not of the edges. These relations appear to be most prevalent adjacent to the quartz-feldspar veinlets and impregnations. In other words it is not always possible to distinguish clearly between textures due to recrystallization in the solid state, and textures formed by the solidification of a melt, although definite and unquestion-

able examples of each can be observed. This is to be expected when one considers the mode of origin of the rocks in this zone; in the pulverization of a hot crystal mesh containing some interstitial liquid it would be inevitable that as the hot crush powder recrystallized it would receive additions from the magmatic materials surrounding it. Thus every transition might be expected to occur between rocks showing cataclastic, crystalloblastic, and hypidiomorphic granular textures.

Field relations indicate that the more mafic the original material the more likely it is to recrystallize. True mylonites formed from the melanocratic inclusions within the migmatite zone are very rare, but are found locally immediately at the contact. Most of the mafic material has been thoroughly recrystallized to granulites and granulitic gneisses of medium to coarse grain. These rocks superficially resemble diorites, but thin sections reveal crystalloblastic textures (Pl. 7, fig. 2) and on definite order of crystallization. They are commonly interlayered with rocks of more felsic composition which are not nearly so extensively recrystallized and may be in the form of mylonites or mylonite gneisses. These mafic granulites and granulite gneisses vary considerably in composition, but the variations apparently are due largely to the extent to which the original ferromagnesian material was impregnated with pegmatitic substances before and during the process of crushing and recrystallization. In the least contaminated types hornblende and plagioclase (andesine or oligoclase) are the dominant minerals. Others minerals commonly sparingly present are pyroxene (diopside), quartz, biotite, sphene, magnetite, and epidote or clinozoisite. From this type of rock gradations in composition may be observed by the entry of potash feldspar (microcline or orthoclase) and the increase in biotite and quartz. In some specimens the quartz and feldspar enter as definite veinlets and impregnations whose boundaries, though welded, are nevertheless distinct. In other cases the material has acted more pervasively and its introduced nature can only be surmised. In extreme cases the introduction of potash has led to the formation of rocks resembling mesocratic syenites and monzonites, but with crystalloblastic textures instead of the textures of igneous rocks. Quartz-rich types in which there has been little addition of potash are also found.

These mafic granulites and granulitic gneisses are commonly disposed in bands of variable thickness separated by blastomylonites or granulitic gneisses of more felsic character into which they grade, or more commonly show welded but fairly abrupt contacts. The felsic material, though commonly less thoroughly recrystallized, usually contains proportionally more of the quartz feldspar mesostasis and as this liquid

fraction increases the rocks grade into definitely igneous types. Furthermore with increase of the feldspar-quartz matrix the felsic layers manifest greater and greater mobility and commonly show complicated drag

TABLE 1.—*Chemical analyses*

	I	II	III	IV	V
SiO_2	67.77	72.46	68.65	57.80	55.26
Al_2O_3	17.69	15.04	15.34	17.46	18.58
Fe_2O_3	Tr	.57	.87	1.33	1.41
FeO	2.79	1.15	2.49	5.94	5.15
TiO_2	.48	.24	.38	.88	.87
MnO	.13	Tr	.07	.11	.06
CaO	3.16	2.71	3.63	7.96	7.69
MgO	.58	.39	1.34	3.87	4.48
K_2O	1.98	2.13	2.24	.67	1.32
Na_2O	4.38	4.70	4.44	2.86	3.83
H_2O-	.10	.19	.18	.15	.05
H_2O+	.62	.30	.20	.67	.78
CO_2	nil	nil	nil	nil	nil
P_2O_5	.16	.09	.22	.22	.44
BaO	ND	.07	.08	ND	ND
SrO	ND	Tr	.04	ND	ND
Totals	99.84	100.04	100.17	99.92	99.92
	Norms				
Quartz	25.4	30.2	23.5	13.8	4.0
Orthoclase	11.7	12.8	13.3	3.9	7.8
Albite	37.1	39.8	37.7	24.1	32.5
Anorthite	14.5	12.8	15.3	32.8	29.5
Corundum	3.1	.2	.....	.....	.....
Dps $CaSiO_3$	.....	.....	.7	2.2	2.4
Dps $FeSiO_3$	.....	.....	.4	1.1	.9
Dps $MgSiO_3$	.....	.....	.3	1.1	1.4
Hyp $FeSiO_3$	4.6	1.2	3.1	7.5	6.1
Hyp $MgSiO_3$	1.5	1.0	3.0	8.6	9.8
Magnetite	.....	.9	1.1	1.9	2.1
Ilmenite	.9	.5	.7	1.7	1.7
Apatite	.4	.2	.5	.5	1.0
H_2O, BaO, etc	.7	.6	.5	.8	.8
Totals	99.9	100.2	100.1	100.0	100.0

I Biotite granodiorite; homophanous facies, core of Mount Annie.
II Swirled gneiss facies, average of 21 samples.
III Gneissic and mylonitic facies, average of 25 samples.
IV Hornblende granulite; mafic layer in gneissic and mylonitic facies.
V Quartz diorite; mesocratic type from homophanous facies, core of Tunk Mountain.
(Analyses by F. Herdsman, Glasgow.)

folding between the adjacent more competent bands of mafic granulitic gneiss. Thus the normally straight foliated rocks of the outer gneissic zone of the batholith pass into the structurally highly complicated zone of swirled and porphyritic gneisses.

CHEMICAL ANALYSES

Available analyses of rocks from the Colville batholith are collected in Table 1. The first three columns give a fair idea of the composition

of this part of the batholith. They are not far from the composition of average granodiorite, except that the ratio of soda to potash is rather high. Column IV is an analysis of a hornblende granulite, probably derived by magmatic permeation and thorough recrystallization of a greenstone xenolith. The last column is an analysis of a ferromagnesian-

TABLE 2.—*Modes (Rosiwal method)*
(in percentages)

	A	B	C	D
Quartz	34	31	18	11
Potash feldspar*	22	7	..	2
Plagioclase	35	50	43	48
Biotite	7	11	10	16
Hornblende	..	..	27	21
Magnetite	1	..	2	2
Muscovite	1	1	..	..

A Biotite granodiorite; homophanous facies, core of Mount Hull.
B Biotite granodiorite; homophanous facies, core of Mount Annie; corresponds to Analysis I, Table 1.
C Hornblende granulite; mafic layer in gneissic and mylonitic facies; corresponds to Analysis IV, Table 1.
D Quartz diorite; mesocratic type from homophanous facies, core of Tunk Mountain; corresponds to Analysis V, Table 1.

* Potash-feldspar includes orthoclase, microcline, and microperthite.

rich type representative of only a small part of the homophanous facies of the batholith.

Rosiwal analyses are given in Table 2. Three of these correspond to Analyses I, IV, and V of Table 1, and the fourth is an analysis of a homophanous granodiorite from Mount Hull for which no chemical analysis was made.

PROTOCLASTIC VERSUS DYNAMIC METAMORPHIC ORIGIN OF THE BORDER ZONE

General statement.—The remarkable structural features in the border zone of the Colville batholith have been explained as the result of a rise of the mass at a time when it was largely crystallized but still contained some residual liquid between the growing crystals. A possible alternative hypothesis is that structures in the border zone are due to dynamic metamorphism acting long after the batholith had solidified. In the following paragraphs arguments against the hypothesis of dynamic metamorphism will be listed, not as an attempt to analyze completely the general problem of foliated border facies, but rather as a means of summarizing our principal conclusions regarding the border zone of this particular batholith.

Evidence from wall rocks.—Dynamic metamorphism of sufficient intensity to produce foliation in the broad outer zone of the Colville

batholith could hardly fail to have affected adjacent Anarchist rocks. The Anarchist series has indeed been subjected to dynamic metamorphism, but several lines of evidence show that this metamorphism preceded and is without direct relation to the deformation of the batholith: (1) The general northwest-southeast trend of regional metamorphic structures in the Anarchist rocks shows little relation to the trend either of the batholithic contact or of foliation within the batholith. The intrusive cuts discordantly across Anarchist folds and foliation, and gneissic structures in the intrusive generally parallel its borders (Pls. 1, 2). (2) Foliation in the batholith bends around projections of the wall rock, and around an Anarchist roof pendant. (3) Inclusions of already metamorphosed Anarchist fragments are occasionally found in the batholith, especially in its pegmatites. (4) Already metamorphosed Anarchist rocks along the contact have been broken into a friction breccia by the stresses which foliated the border gneiss.

The fact that deformation of the batholith was accompanied only by the fragmenting of Anarchist rocks at its immediate border must mean that the phyllites, greenstones, and chlorite schists of the Anarchist series were actually stronger and more brittle than the border zone of the intrusive. This condition could hardly have existed if the batholith had been solid at the time of deformation. The evident difference in strength between wall rocks and intrusive can be accounted for only by the assumption that the batholithic border consisted of a crystal mesh with a residuum of still uncrystallized magma.

Evidence from batholithic border.—Several features observed within the border zone of the batholith also afford cogent reasons for rejecting the alternative hypothesis of dynamic metamorphism: (1) Regional control of the foliation in the batholithic border rocks might reasonably be expected as a result of dynamic metamorphism. Such regional control is almost wholly lacking; the foliation turns curves controlled by irregularities in the wall, forms domes and basins, commonly crosses the schistosity of Anarchist rocks at high angles. Only the northwest-southeast linear structure of the Colville mass has a possible regional significance.

(2) Deformation is most extreme near the outer margin of the intrusive, decreasing steadily from the zone of banded and mylonitic gneiss through the swirled gneiss zone to the homophanous zone. On the hypothesis of dynamic metamorphism, such a transition from granitic rocks at the core to strongly foliated rocks at the periphery would be possible provided that the margin of the mass was its most plastic portion during metamorphism. This would imply that not only shear but recrystalliza-

tion and evidence of plasticity should decrease toward the interior of the mass. A quite different sort of transition is apparent in the Colville batholith—between the strongly foliated rocks at the periphery and the homophanous rocks of the interior is a zone whose intricate swirling and involved convolutions imply greater instead of less plasticity than the outer zone. Furthermore, recrystallization increases steadily from the margin through the zone of swirled gneiss until the recrystallized minerals give way to those of definitely magmatic origin. An increase in plasticity and degree of recrystallization away from the zone of most pronounced shearing is hardly possible unless the interior of the mass was more fluid than the exterior at the time of deformation.

(3) The quartz-feldspar mesostasis which pervades the border rocks, penetrating into crushed and deformed crystals simultaneously with the deforming movements, and locally completely squeezed out during deformation to collect in pools at the periphery, can be explained on the hypothesis of dynamic metamorphism only as a liquid injected from an extraneous source during metamorphism. Perhaps, according to the hypothesis of dynamic metamorphism, concentration of the pegmatitic material at the border of the batholith might be expected, since this would be the zone of greatest deformation and of easiest escape for fluids; but it seems odd that no pegmatite should occur as discrete bodies in either Anarchist rocks or the batholith itself at points far from the contact. Even more troublesome to explain is the gradual transition of the mesostasis into first the igneous groundmass of rocks in the swirled gneiss facies and finally into the uniform igneous rocks of the homophanous facies. Relations of the mesostasis to the contact and to the homophanous granodiorite, so difficult to account for on the assumption of dynamic metamorphism, receive a ready explanation from the hypothesis of protoclastic deformation.

(4) A considerable amount of structural evidence indicates that the Colville mass during its deformation was rising with respect to a more stationary wall: the direction of rolling of feldspar augen; the pattern of faults and the directions of slickensiding on fault surfaces both in the zone of friction breccia beyond the batholith and within the border of the batholith itself; the orientation of the axial planes of swirls and folds in the zone of swirled gneisses; the displacement of pegmatite dikes which filled cross joints and were then made into steplike dikelets by a continuation of foliation plane slip after the cross joint was formed. While suitably oriented stresses from an external source might conceivably produce such structures, they are far more simply explained by upward movement of the Colville mass itself.

APPLICATION TO THE GENERAL PROBLEM OF THE INTERPRETATION OF FOLIATED BORDER ZONES

A brief comparison of the Colville border zone with other areas of similar structure may be of interest.

Structures somewhat analogous but by no means identical with those of the Colville mass have been described in the igneous rocks of the Adirondacks. Balk (1931) considered these structures to be of protoclastic origin, but more recently Buddington (1940) has ascribed them to dynamic metamorphism. In at least one important respect conditions in the Adirondacks are very different from those in the Okanogan Valley,—the country rocks are highly metamorphosed, suggesting that their temperature may have been high when the igneous body was intruded. This circumstance makes inapplicable most of the criteria previously discussed. Had the temperatures of the wall rocks into which the Colville batholith rose been higher, widespread regional recrystallization might well have wiped out all evidence of the protoclastic manner of emplacement of the mass.

The similarity of the swirled and migmatitic gneisses of the Colville border zone to the Archean gneisses of Canada and Scandinavia is apparent from the detailed descriptions of many Archean districts. Whether the banding in these Archean gneisses is a flow structure or a metamorphic structure is an old geologic problem about which much has been written. In these rocks, as in those of the Adirondacks, the problem is nearly always complicated by a higher degree of metamorphism of the intruded rocks and by more extensive recrystallization than found in the Okanogan Valley. Despite these differences, the marked resemblance in structural detail,—gradual transitions from foliated to unfoliated rocks, intense swirling, banded gneisses passing into migmatites, and the intimate association of unfoliated and slightly foliated pegmatites with highly foliated gneisses—suggest that many Archean gneisses may be of protoclastic origin. Wegmann (1930) and others have pointed out the similarity between the structural pattern of some ancient intrusives and that of salt domes, and have suggested that these intrusives rose in diaperlike fashion. Movement of an intrusive in the last stages of its solidification should not, seemingly, be a highly unusual phenomenon, and such movement under differing conditions of temperature and pressure might well produce structures like those of the Colville batholith on the one hand and those of Archean gneisses on the other. If the palingenetic origin of granitic magmas is accepted there is a double opportunity for the deforming stresses to have the correct temporal relations to bring about the characteristic structural details inasmuch as the requisite degree of plasticity would be achieved not only

during the last stages of consolidation, but also in the early stages of palingenesis when ultrametamorphism had advanced to the point where a tenuous ichor of interstitial fluid was beginning to form throughout the rock. If protoclastic deformation occurred during the early stages of palingenesis the extremely high temperature of the wall rocks would result in the erasing of the record by recrystallization even more commonly than in those cases where protoclastic rise occurred during the last stages of solidification. Quite possibly the Colville mass, because of the low temperature of its wall rocks and the absence of later extensive deformation, has fortuitously preserved the evidence of a sequence of events which was common in many ancient intrusives.

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STRATIGRAPHIC TERMINOLOGY

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ABSTRACT

In order to attain uniformity in stratigraphic terminology it is here proposed that a distinction be made between lithogenetic, time-stratigraphic, and time units. A three-fold arrangement of appropriate terms is given in tabular form.

INTRODUCTION

The classification of strata is amenable to a logical, orderly, and at the same time flexible arrangement. It follows that uniform terminology can be made a practical tool in stratigraphic geology. Presented here is a greatly condensed treatment of a general principle which underlies the application of stratigraphic terms; it is a principle which accords with the procedure of the field geologist. The writers believe that the application of these ideas will lead ultimately to a more uniform and stable stratigraphic terminology, as the arrangement is sufficiently elastic to serve all the needs of geology.

LITHOGENETIC UNITS

The fundamental need is a system of terminology that will permit the geologist to define the lithogenetic units as he finds them in the field and in subsurface sections, without bias concerning their ages. He delineates these concrete rock units objectively on his map and sections. Other geologists may at any time go into the area and find the rocks

he named and mapped. This is by no means a new idea. In 1888 Powell made the following statement:

> "The maps are designed not so much for the specialist as for the people, who justly look to the official geologist for a classification, nomenclature, and system of convention so simple and expressive as to render his work immediately available alike to the theoretic physicist or astronomer, the practical engineer or miner, and the skilled agriculturist or artisan. * * * Accordingly, the classification involved in a cartographic system designed for general use should be objective rather than theoretic; it should be based upon rock-masses in their observed and readily observable relations rather than upon time-intervals contemplated in historic geology or even upon the organic remains contemplated in biotic geology; it should be petrographic rather than chronologic or paleontologic. * * * But while the minor geologic divisions must therefore have a natural basis, those of greater magnitude may be somewhat differently defined. * * * The structural geologic unit is the "formation". It is defined primarily by petrography and secondarily by paleontology; and, in thoroughly studied regions, is generally found to constitute a genetic unit."

The geologist mapping a given region rarely deals with a complete section. He rarely finds it possible or relevant to his problem to show the exact relations of the cartographic units in his area to similar units in Europe, Asia, and Africa, or even to units in less distant regions. Even more rarely can he use fossils to recognize paleontologically defined subdivisions of the formations he has mapped. Such matters as these he leaves to the biostratigrapher.

TIME-STRATIGRAPHIC UNITS

Next, then, comes the task of the biostratigrapher. He is a stratigrapher who classifies strata by biologic criteria. He is a co-ordinator and synthesizer of stratigraphic data accumulated by many field geologists. The evidence employed by the biostratigrapher is usually complex, and he may draw upon data from far-removed localities. One of his main tools is the standard section of reference of inter-regional or regional scope—a composite section assembled from the most complete sequences of strata at several localities. Consequently, the terminology of the biostratigrapher is not apt to coincide in every respect with that of the cartographer.

Further, the biostratigrapher deals with units of a standard column which are delimited by fossils regardless of kind of rock and thickness of strata. Since the fossils usually serve as our nearest approach to time-markers, it follows that the units of this standard column are stratal units defined by time. This is in contrast with the units dealt with in mapping formations, as these are controlled primarily by lithology. The biostratigrapher strives to correlate the local mappable units with those of the standard scale in order that the physical and biologic phenomena read from a local study of the formations may be fitted into the general sequence of geologic events of a wider region.

TIME TERMINOLOGY

The third step in stratigraphy geology is thus to present a sequence of events. This is done in terms of geologic time of the standard chronologic scale. This scale is composed of purely time terms, such as Era,

TABLE 1.—*Stratigraphic terminology (simplified)**

Time		Rock
I Time terms	II Time-rock terms (Time-stratigraphic)	III Lithogenetic terms for units of more or less local extent
1. Era *e. g.* (Mesozoic Era)	1. No specific term is recommended. Rocks or strata in a general sense will answer the purpose,* *e. g.*, Mesozoic rocks.	Group Formation Member, lentil, tongue Bed, stratum, layer
2. Period	2. System (Triassic System)	
3. Epoch	3. Series (Upper Triassic Series)	
4. Age	4. Stage	
	5. Zone	

* If the time and time-rock terms are used in a restricted technical sense, according to the above scheme, they may be capitalized, *e.g.*, Triassic Period.

Period, Epoch, and Age. It would not only be awkward but actually incorrect to say, for example, that certain dinosaurs lived in the Upper Jurassic Series—they lived during the Late Jurassic Epoch.

THREE CATEGORIES OF TERMS

The procedure discussed above necessitates the adoption of three categories of terms: (1) time, (2) time-stratigraphic, and (3) lithogenetic or cartographic. In order that the differences and relations between these may be clearly understood an arrangement of the most fundamental units is shown in Table 1. In order that the reader may compare this arrangement with the schemes followed by other authors, a compilation from several standard references is presented in Table 2.

TERMS FOR LITHOGENETIC UNITS

Column III of Table 1 includes the terms applied to the rocks which the geologist maps, describes, samples, and photographs. The funda-

TABLE 2.—*Comparison of usage of stratigraphic terms* *

I. G. C. Bologna, 1881		J. D. Dana, 1895		I. G. C. Paris, 1900 (Renevier, 1901)		Lapparent, 1900		Geikie, A., 1903		Blackwelder and Barrows, 1911; L. C. Snider, 1932	
ERA	Group	AEON	Series	ERA	....	ERA	Group Series	ERA,		ERA	Group
PERIOD	System	ERA	System	PERIOD	System	PERIOD	System	PERIOD	System	PERIOD	System
								EPOCH, AGE†			
EPOCH	Series	PERIOD	Group	EPOCH	Series	EPOCH	Stage	"(?)	Series, Section, Formation	EPOCH	Series
AGE	Stage	EPOCH	Stage	AGE	Stage	——	Substage	"(?)	Group or Stage		
——	Assise			PHASE	Zone	Age	Assise	"(?)	Beds or Assise		
								"(?)	Zone or Horizon		
								"(?)	bed lamina		
								† . . . "terms . . . equivalent, or at least interchangeable" . . .			

H. F. Cleland, 1916 E. Kayser, 1923 W. J. Miller, 1929 Gregory and Barrett, 1931		W. B. Scott, 1932		R. C. Moore, 1933		Schuchert and Dunbar, 1933		U.S.G.S., I.G.C. Guidebook, 1933 (Reeside)		A. H. Sutton, 1940	
ERA	Group	ERA	Group	ERA	Sequence			ERA	——	ERA	Sequence
PERIOD	System	PERIOD	System	PERIOD	System	PERIOD	System	PERIOD	System	PERIOD	System
EPOCH	Series	EPOCH	Series	EPOCH	Series	EPOCH	Series	EPOCH	Series	EPOCH	Series
AGE	Stage	AGE	Stage	AGE	Group	STAGE	Formation	EPOCH	Group	SUB-EPOCH	Sub-series
		HEMERA	Substage	STAGE	Formation	——		EPOCH	Formation	CHRON	Group
		——	Zone	SUB-STAGE	Member			EPOCH	Member	STAGE	Formation
										SUB-STAGE	Member

* For another compilation of terms, consult Kruizinga (1929).

mental unit in this category is the formation, which may be subdivided lithologically into members and smaller divisions. The formation thus mapped does not carry at the outset any connotation of time until an attempt is made to ascertain the time interval during which it was deposited. This time is determined by the use of paleontologic evidence. It is at this moment that the geologist resorts to a reference time scale and may find that a formation he had just mapped represents a fraction of some division in the standard scale, such as a Series, or it may be of the magnitude of a System. Moreover, it may be at least in part of a different age from place to place, although many formations are of the same age over wide areas. A formation represents a deposit of sediments[1] accumulated under relatively uniform conditions; it is formed of more or less similar rock constituents. A formation is therefore a lithogenetic unit. The conclusion is that the terms listed in column III are strictly objective and pertain to the concrete entities mapped by the geologist or recognized by him in subsurface sections. In order to focus attention upon the fact that the lithogenetic units do not necessarily bear any direct relation to the units in columns I and II, they are set off and written in such a form as to preclude any fixed notion of a correlation.

TERMS FOR TIME-STRATIGRAPHIC UNITS

The terms in column II of Table 1 are the stratal terms delimited by time. This time-stratigraphic category comprises material stratigraphic units consisting of sediments deposited during a given time interval; they have time boundaries only, and are independent of lithologic facies. No regard is here paid to thicknesses of strata. The boundaries between these units are established primarily on a paleontologic basis and, as a result, are to a considerable degree subjective. These boundaries are more or less arbitrarily fixed. Unlike the lithogenetic units which are of local extent, the units in column II are of regional or even interregional application. As the units in both columns II and III refer to rocks, they are both rock terms. But in column II these rock terms are defined by time; hence, they are referred to as time-rock or time-stratigraphic units.

TERMS FOR TIME UNITS

The terms in column I of Table 1 are for time units. They are directly comparable with those in column II and are designated by the same proper names, such as Devonian System and Devonian Period. One

[1] Igneous and metamorphic rocks are not considered here.

* For another compilation of terms, consult Kruizinga (1929).

cannot define time terms without the corresponding time-stratigraphic terms.

TIME-STRATIGRAPHIC AND LITHOGENETIC UNITS IN CONTRAST

It is beyond the scope of this paper to analyze the history of the establishment of the existing Zones, Stages, Series, and Systems. The writers do not propose to review the rise of stratigraphic nomenclature but are limiting themselves to a brief survey of the main characteristics of the time-stratigraphic units as they apply to the present scheme. This is perhaps best explained by sketching the steps followed by the stratigrapher in setting up such units.

First, the stratigrapher carefully studies a continuous section of similar facies; collections of fossils are carefully allocated in this section. Second, the species are identified, and their stratigraphic ranges in this section are accurately determined. Third, these ranges are analyzed so as to show a certain grouping of the strata; these are the arbitrarily delimited, provisional, time-stratigraphic units. Each such unit contains species or genera restricted to it; some fossils have their lowest stratigraphic occurrences in it; and some may occur above and below as well as in the unit. Other species which occur first in the immediately overlying beds range no higher than the superjacent unit. Fourth, these time-stratigraphic units are tested by determining a similar distribution of fossils at some other more or less distant locality. Repeated testing proves the validity of the units for the entire geologic province. The sequence of time-stratigraphic units established on such a basis would serve as a yardstick, and its application to more distant areas may be justified if supported by valid paleontologic evidence.

Unconformities—gaps in the geologic record—have no place in such a scheme. Continuous sections, preferably of neritic deposits, offer the best possibility of control. Relative magnitude or scale determines whether these biostratigraphic units are to be treated as Zones, Stages, or some higher category. By controlled piecing together, these units may be connected with stratigraphically adjacent units of some other region. The final result is an uninterrupted succession of stratal units delimited by paleontologic evidence, or, in other words, by time.

In contrast, the lithogenetic units—formations or units of other magnitude—are delimited by changes in lithologic character and are often demarcated by unconformities. Here the character of the rock plays the most important part in the definition of the unit, whereas a consideration of lithology did not constitute the basis for the definition of the time-stratigraphic units. Fossils in lithogenetic units play a secondary role, as they are merely distinctive constituents of the rock. These

fossils, of course, provide a means of correlating a formation with the standard time-stratigraphic scale.

CONCLUSION

In summary, the foregoing discussion was designed to show that the practical needs of geology require three categories of terms: first, the lithogenetic or cartographic terms, which are to apply to the local mappable units; second, the time-stratigraphic terms which are applied to the divisions of the standard scale of reference—standard column—of regional or interregional scope; and the third, purely time terms. The interrelation and ranking of the most fundamental terms given in Table 1 summarize this three-fold arrangement.

The writers are indebted to many colleagues for their suggestions and constructive criticisms during the preparation of this paper.

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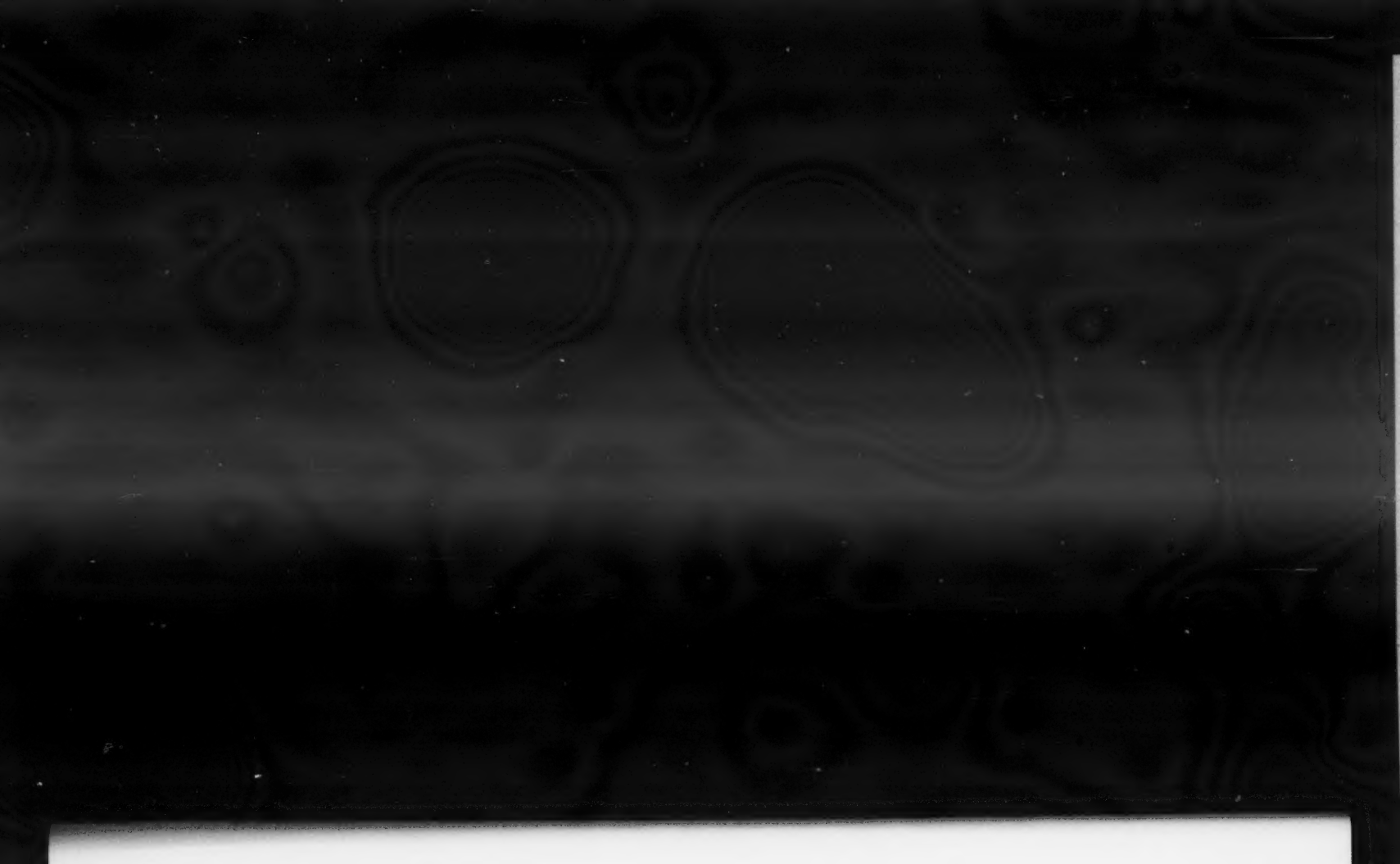

TABLE 1.—*Standard of the Uppermost Jurassic and zo*

STAGES & SUBSTAGES		ZONES & SUBZONES		ZONE Aff
TITHONIAN	AQUILONIAN	Berriasella privasensis	Berriasella delphinensis Protacanthodiscus chaperi	
	BONONIAN	Titanites giganteus	Torquatisphinctes tenuistricta Proniceras idoceroides	Titanite Crendoni
		Provirgatites scythicus		Proglaba Holocepha
		Subplanites contiguus	Pavlovia rotunda Virgatosphinctes pseudoulmensis	Pavlovia Pectinat Virgatos Subplani
		Neochetoceras steraspis	Subplanites virgineus Waagenia hybonota Waagenia beckeri	Subplani Gravesia Gravesia
KIMMERIDGIAN		Aulacostephanus eudoxus	Sutneria subeumela Aulacostephanus pseudomutabilis	Aulacost
		Streblites tenuilobatus	Idoceras balderum Ataxioceras polyplocum	Rasenia
		Sutneria platynota	Sutneria platynota Idoceras planulum	Rasenia Pictonia
OXFORDIAN		Peltoceras bimammatum		Ringsteadi
		Aspidoceras hypselum		Virgatosp
		Peltoceras transversarium		Amoebocer
		Aspidoceras perarmatum		Cardioce
		Cardioceras cordatum		Cardioce
		Quenstedioceras mariae		Quenstedi
		Quenstedioceras Lamberti		Quenstedi

sic and zonal sequences in England, southeast Europe, and Russia

	ZONES IN ENGLAND & After L. F. Spath	NORTHWEST EUROPE After W. J. Arkell	ZONES IN SOUTHEAST EUROPE	ZONES IN RUSSIA ("BOREAL" PROVINCE)
	(Purbeck beds)	(Purbeck beds)	Berriasella maljoi Berriasella ciliata	Riasanites rjasanensis Craspedites nodiger Craspedites subditus Craspedites okensis Kachpurites fulgens
	Titanites giganteus Crendonites gorei Proglabanites vulgaris Holcophinctes pallasioides Pavlovia rotunda Pectinatites pectinatus Virgatosphinctes wheatleyensis Subplanites grandis Subplanites virgatus Gravesia irius Gravesia gravesiana	Titanites giganteus Kerberites kerberus Crendonites gorei Provirgatites scythicus Pavlovia rotunda Pectinatites pectinatus Subplanites sp. Gravesia irius Gravesia gravesiana	Oppelia lithographica Waagenia hybonota Waagenia beckeri	Epivirgatites Nikitini Virgatites virgatus Provirgatites scythicus Provirgatites niatschkowiensis Subdichotomoceras bleicheri
	Aulacostephanus pseudomutabilis Rasenia mutabilis Rasenia cymodoce Pictonia baylei	Aulacostephanus pseudomutabilis Rasenia mutabilis Rasenia cymodoce Pictonia baylei	Sutneria subcumela Aulacostephanus pseudomutabilis Idoceras balderum Coccotraustes dentatus Ataxioceras suberinum Sutneria platynota Sutneria galar Idoceras planulum	Aulacostephanus eudoxus Streblites tenuilobatus
	Ringsteadia pseudocordata Virgatosphinctes decipiens Amoeboceras alternans Cardioceras cawtonense Cardioceras cordatum Quenstedioceras mariae Quenstedioceras lamberti	Ringsteadia pseudocordata ? Perisphinctes variocostatus Perisphinctes plicatilis Cardioceras cordatum Quenstedioceras mariae Quenstedioceras lamberti	Peltoceras bimammatum Peltoceras uhligi Peltoceras transversarium Aspidoceras perarmatum Cardioceras cordatum Quenstedioceras mariae Quenstedioceras lamberti	Primodoceras serratum Amoeboceras alternans Cardioceras vagum Cardioceras cordatum Quenstedioceras mariae Quenstedioceras lamberti

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STANDARD OF THE JURASSIC SYSTEM

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ABSTRACT

The conflicting use of terms pertaining to the Standard of the Jurassic System is briefly reviewed, and an attempt is made to reconcile these controversial issues. The recommended Standard of the Jurassic, representing a composite section of local but most completely developed fossiliferous sections from different parts of the Old World, is tabulated on the left side of Table 2. It is to be hoped that the use of this recommended Standard will contribute to a greater precision of stratigraphic expressions and will result in a more uniform terminology.

INTRODUCTION

Stratigraphic investigations of Mesozoic rocks, particularly of the Jurassic, conclusively demonstrate the feasibility of a more or less precise correlation and accurate dating of rocks, in terms of the European

Standard column. For many years stratigraphers, engaged in the study of rocks other than Jurassic, have looked with envy on the refinement that has been achieved in the classification of Jurassic strata and on the success with which the Jurassic Standard has been applied in widely separated regions. Unfortunately, however, this apparent achievement in accuracy and precision is appreciably offset by the ambiguity of terminology arising from the unnecessary multiplicity of terms and the lack of uniformity in the application of these terms to the various Standard divisions.

The widely divergent interpretations of the scope of series, stages, and substages compel a writer on Jurassic stratigraphy either to give his own definitions of terms or to refer the reader to some previous authoritative publication in which precise definitions of the terms may be found.

Today there are in use almost as many different "Standards" of the Jurassic System as there are comprehensive treatises on the subject. The compilation on Table 2 shows only a few of these so-called "Standards", many of which at one time or another had been accepted by a student of Jurassic stratigraphy as an authoritative Standard column.

The divergence of views regarding the subdivisions of the Jurassic System appears to have existed in the literature since the days of Quenstedt and Oppel. It will be remembered, for example, that Quenstedt, following Buch, had placed the upper limit of the Middle Jurassic Series above the strata now generally classified as the Callovian Stage, whereas Oppel, in his monumental work on the Jurassic System, drew the boundary between the Middle and Upper Jurassic below the Callovian Stage. (*See* Table 2.)

Today hardly a subdivision of the Jurassic System is free of controversy as to its exact scope and nomenclature. There is even no general agreement regarding the limits of the Jurassic System. Some French geologists still hold the view that the Jurassic System begins at the base with the Rhaetian Stage, whereas outside of France the Rhaetian Stage is regarded almost universally as the uppermost division of the Triassic System. Other French geologists go to another extreme and restrict the name Jurassic to the rocks below the Cretaceous and above the Lias. Lias is regarded by these men as an independent system.

The difference of opinion as to what constitutes the Middle Lias is another striking example. Continental geologists place in the Middle Lias the entire Pliensbachian Stage, whereas to most British geologists the Middle Lias includes only the upper part of the Pliensbachian Stage —the Domerian Substage.

As can be seen from Table 2 the controversial usages of the Stage terms are very numerous. In the subsequent pages attention will be given only to those cases which are pertinent to the Standard of the Jurassic proposed in this article.

PURPOSE AND PLAN

As there is no individual authority nor a group of men vested with mandatory power to enforce the use of one or the other scheme of stratigraphic terminology, the validity and general acceptance of terms are determined to a considerable degree by a widespread voluntary application or usage of these terms by individual authors.

It is thus only through a co-operative effort and a general agreement on terms that a much-desired uniformity as well as greater precision and accuracy of expression can be achieved. With this aim in mind the writer appeals to stratigraphers, and particularly to those who are engaged in the study of the Jurassic, to analyze critically the existing chaotic state of the terminology and to give thorough consideration to the scheme recommended in this article as the Standard of the Jurassic System.

It should be borne in mind that in regard to some parts of the Jurassic System opinions are at present fairly evenly divided, and future research might necessitate slight changes of the Table as here proposed.

An exchange of views on such issues through press or correspondence is invited, and the writer will be only too glad to volunteer his services as custodian of such documents for a subsequent revision of the Standard and, perhaps, for a presentation of the entire case at some future date to the International Geological Congress for its formal approval and recommendation.

ACKNOWLEDGMENTS

The critical study of the component units and terminology of the Jurassic Standard was undertaken in connection with the investigation of the Lower Jurassic stratigraphy and paleontology of west-central Nevada (Muller and Ferguson, 1939).

Further encouragement to carry this problem to completion was received from H. G. Schenck, who, while engaged in exploration of the little-known terrain of Iran for the Amiranian Oil Company, asked the writer to send him a Standard table for the entire Jurassic System. The experience of field men in Iran, whose interpretation of geologic structures depended upon the stage designation of the strata, provides a vivid illustration of practical implications that are likely to arise

from the application of a conflicting terminology. For instance, designation of the two adjacent sets of strata as "Lower Bajocian" and "Upper Aalenian" permits no less than three different interpretations of their structural relationships—depending upon whose "Standard" of reference is accepted by the field geologist. The strata in question may be taken to mean a normal, upright sequence of beds; they may be interpreted as representing an overturned section, or even an overthrust.

While carrying on the present investigation of the controversial terminology of the Jurassic System the writer had a rare opportunity to discuss the problem with many leading stratigraphers in England and central Europe. He also has examined and thus obtained firsthand knowledge of a number of Jurassic sections in southern England and Germany. For this opportunity to study abroad the writer is indebted to the Geological Society of America whose generous grant-in-aid from the Penrose Bequest is here gratefully acknowledged.

The writer also wishes to express his indebtedness and deep gratitude to his many European colleagues for their inspiring encouragement and for their helpful suggestions and constructive criticism of the hereinproposed Standard. These men have unhesitatingly placed at the writer's disposal their intimate knowledge of the Jurassic and have been of invaluable aid in the difficult task of reconciling the existing controversial issues. The writer wishes to thank particularly Dr. W. J. Arkell of the Oxford University, Dr. Edgar Dacqué of the Institute in Munich, Dr. W. O. Dietrich of the Institute in Berlin, Dr. Paul Dorn of the University of Tübingen, Dr. Edwin Hennig of the University of Tübingen, Dr. Erich Jaworski of the University of Bonn, Dr. Oskar Kuhn of the University of Halle, Dr. Julius von Pia of the Vienna Museum of Natural History, Dr. Joachim Schröder of the Institute in Munich, Dr. L. F. Spath of the British Museum (Natural History), Dr. Friedrich Trauth of the Vienna Museum of Natural History, and Dr. T. H. Withers of the British Museum (Natural History).

Thanks are also due other members of the staffs of the museums, universities, and libraries visited during his study abroad. At all these places the writer enjoyed the warmest hospitality and most courteous assistance.

GUIDING PRINCIPLES FOR THE SELECTION OF UNITS FOR THE RECOMMENDED JURASSIC STANDARD

The task of reconciling the existing differences as to the scope and terminology of the various units of the Jurassic Standard is by no

means a simple one, and it is only natural to expect that the scheme recommended here may not be approved in its entirety by some workers. Several borderline issues still remain on which no definite recommendation has been made at this time.

In contrast with the wide divergence of views among the authorities on the units of higher rank, there appears to be a more general agreement on the succession of faunal zones and subzones. These zones and subzones, based on the ammonite assemblages, provide the only workable means for determining the stratigraphic limits of the larger units in the Jurassic Standard as used by the different authors. It is fully realized that the compiled table of zones and particularly of subzones, based on local sequences in different parts of the Old World, may not be applicable in its entirety or in its every detail throughout the globe. It is hoped, however, that this detailed sequence will provide a convenient means for testing this very fact. It thus furnishes a working scheme—something to test by further investigation.

In view of the above considerations the proposed Standard of the Jurassic System should be regarded only as a provisional arrangement, subject to further corrections and additions.

In stratigraphy, no hard and fast rules exist to aid one in determining which of the several stage terms is to be regarded as valid. It follows from general practice, however, that well-established usage should be given considerable if not deciding weight. Accordingly, in the proposed Standard of the Jurassic System, usage has been given primary consideration except where it was found that the application of the term was contrary to other fundamental principles of procedure in stratigraphy.

In most cases the application of the usage criterion is not difficult, but where two or more terms had been equally widely used for the same stratigraphic unit other factors had to be considered. In such cases, and when other conditions were more or less equal, the validity of the term was decided on the basis of priority.

Arkell in his classical work on the Jurassic System in Great Britain has tabulated in the appendix over 100 terms which at one time or another had been proposed as the stage names for the Jurassic. Many of these terms represent exact synonyms of previously established and widely used names and therefore can be disposed of without further argument. Thus the Charmouthian, proposed by Mayer in 1864 for exactly the same stratigraphic unit which had been previously called Pliensbachian by Oppel, is not included in the list of recommended terms, as it is an exact synonym of the Pliensbachian. But the rule of priority should not be blindly followed, and the reader will agree that a still earlier name for

the same unit, the Liasian, proposed by D'Orbigny, has been justifiably suppressed because of its phonetic similarity to the Liassic (Lower Jurassic) Series.

The lack of uniformity in the use of some stage terms is directly ascribed to the failure of some authors to distinquish the well-established stages (units with fixed time-stratigraphic limits) from the local, formational (lithogenetic) units bearing the same or similar names (Schenck and Muller, 1941). Such cases have been critically scrutinized, and the preferred terms have been chosen in compliance with the principles outlined above. The Kimmeridgian Stage and the Kimmeridge Clay serve as an example. The Kimmeridgian Stage was proposed by D'Orbigny and soon afterward was revised by others to include the strata bounded below and above by the beds characterized by certain faunal assemblages or zones. The Kimmeridgian Stage was thus preceded by the Oxfordian Stage and was followed by the Tithonian Stage, and in this scope the Kimmeridgian Stage attained wide acceptance and had become deeply rooted in the literature. Only a little more than a decade ago additional faunal zones were recognized by some British geologists in the upper part of the Kimmeridge Clay. In the already established Stage classification of the Jurassic System these new faunal zones represented a part of the section above the uppermost limit of the Kimmeridgian Stage. Because of this extended faunal sequence in the Kimmeridge Clay some British stratigraphers have shifted the upper boundary of the Kimmeridgian Stage to include these additional zones, whereas other workers retained the previously established boundary of the Kimmeridgian Stage and placed the newly discovered faunal zones of the Kimmeridge Clay in the next higher, the Tithonian Stage.

The situation is indeed confusing and may well serve as a warning to stratigraphers in the future not to employ the names of local formations for the stages of the standard sequence—especially if there is any dispute as to the stratigraphic boundaries of the formation in question. Tertiary stratigraphers in California, who insist that the formational name Vaqueros is a suitable stage name for the Standard of Miocene or Oligocene, will do well to heed the consequences experienced by the Jurassic men.

It is recommended here that the newly recognized faunal zones in the uppermost Kimmeridge Clay should properly be placed in the Tithonian Stage regardless of their unbroken continuity and lithological affinity with the underlying zones of the Kimmeridgian Stage.

It would have been a different problem had the new faunal zones from the uppermost Kimmeridge Clay represented an altogether new assem-

blage, filling the gap between the Kimmeridgian Stage and the next higher Tithonian Stage. It would have been proper then to extend the limits of the Kimmeridgian Stage to include these new zones.

A discrimination of the faunal facies on which the different stages are based has been applied also in the selection of the stages for the proposed standard. As the present Standard of the Jurassic is based on the ammonite facies, such stages as Corallian and Virgulian are eliminated from the recommended list.

The above procedure, however, does not imply that there is no need in stratigraphy for a Standard based on a different faunal facies. The need for such an auxiliary Standard of reference is realized by many workers, and for some parts of the geologic column it has already been filled. In the Standard of the Carboniferous and Permian systems, for example, separate zonal sequences are recognized, based on foraminifers, cephalopods, corals, brachiopods, plants, and fresh-water mollusks. Such a scheme for the Jurassic and other systems would be desirable and no doubt will be worked out in the future.

Ambiguity and confusion are also introduced into the terminology of the Jurassic System when a part of a well-established stage is designated by a new stage name without any provision for the remaining part of the original stage. An example of such a procedure is the case of the Aalenian Stage. The Aalenian Stage was introduced for the lower part of the Bajocian Stage of the Middle Jurassic with no specific recommendation as to what is to be done with the Bajocian. In the current literature some workers still use the Bajocian in its original sense and see no need for the Aalenian; others employ the name Aalenian as an independent stage and restrict the Bajocian to what remains above the Aalenian; and still others treat the Aalenian as a substage of the Bajocian but leave the upper part of the Bajocian unnamed.

This is one of the cases on which the writer did not feel ready to make a definite recommendation. Because of the evenly divided usages further expressions on this issue are invited. Provisionally the name Aalenian is left out of Table 2 as is shown in the left-hand side of the Table.

The concept of the Standard of the Jurassic Period, divided into "Ages" and hemerae, introduced by Buckman, is such a radical departure from the generally accepted terminology and established principles of stratigraphic chronology that its abandonment is recommended. The merits of Buckman's scheme have been questioned and in part disproved by his British colleagues. Furthermore, it is very undesirable to employ the term "Age" in the new restricted technical sense of Buckman when this term is already widely used as a time term corresponding to the stage.

The instability of the generic names of ammonites after which the "Ages" were named was voiced by some as another objection to Buckman's scheme of chronology.

MAJOR DIVISIONS OF THE JURASSIC SYSTEM

GENERAL STATEMENT

According to prevailing usage the Jurassic System is divided into three main parts or series: Malm or Upper Jurassic, Dogger or Middle Jurassic, and Lias or Lower Jurassic.

LIAS OR LOWER JURASSIC SERIES

General statement.—Practically all geologists define the Lower Jurassic Series or Lias as beginning with the Zone of *Psiloceras planorbis* and ending with the Zone of *Lytoceras jurense*. Only a few men among the French begin the Lower Jurassic Series with the Rhaetian Stage, basing their argument on the fact that the marine rocks of the Jurassic System were deposited during a single major transgression which began in most of western Europe during the Rhaetian Age. This view, however, is not shared by the greater majority of stratigraphers, and in the proposed scheme it is recommended that the base of the Jurassic System be taken to begin with the strata containing the earliest Jurassic ammonite genus *Psiloceras*, above the Rhaetian beds with *Pteria contorta* and *Choristoceras marshi*.

According to Oppel and most of the later authors the Lias is subdivided into the Lower Lias, Middle Lias, and the Upper Lias. The Lower Lias includes the Hettangian and Sinemurian Stage. The Middle Lias corresponds to the Pliensbachian Stage which is subdivided into the Carixian Substage below and the Domerian Substage above. Some authors restrict the Middle Lias to the upper part of the Pliensbachian Stage—the Domerian Substage—but it is recommended that the entire Pliensbachian Stage be included in the Middle Lias in conformity with the usage established by Oppel and accepted by the majority of stratigraphers today, except in England. The Upper Lias is almost unanimously defined as representing the Toarcian Stage.

Hettangian Stage.—This stage was established by Renevier in 1864 for the lower part of D'Orbigny's Sinemurian Stage, and, according to Renevier's original definition, includes the Zones of *Psiloceras planorbis* and *Schlotheimia angulata*. In recent years these two zones have been rather minutely subdivided as shown by Spath (*in* Lang, 1924) and by W. Lange (1924, 1931). The applicability of these faunal subdivisions

(subzones) in interregional correlation still remains to be proved, and they are offered here, therefore, with some reservation.

The as yet unbridged faunal gap between the Uppermost Triassic (Rhaetian) and the Lowermost Jurassic (base of the Hettangian Stage) indicates that there is a likelihood of a downward extension of the Hettangian Stage should a pre-*Psiloceras* faunal sequence be discovered at some future date, unless it should prove to be of sufficient stratigraphic magnitude to warrant introduction of a separate new stage.

Sinemurian Stage.—This stage was originally established by D'Orbigny for the entire Lower Lias up to and including the beds with *Echioceras raricostatum.* In 1864 Renevier proposed the name Hettangian for the lowermost two zones of the original Sinemurian, and from then on the name Sinemurian has been consistently applied by practically all stratigraphers to the strata beginning at the bottom with the Zone of *Ammonites bucklandi* and ending at the top with the Zone of *Echioceras raricostatum.* The subzones of the Sinemurian Stage have been compiled chiefly from the works of Buckman and Spath and are offered here only as a tentative working scheme with no claim for finality.

Pliensbachian Stage.—This stage includes the strata beginning at the bottom with the Zone of *Uptonia jamesoni* and ending at the top with the Zone of *Amaltheus (Paltopleuroceras) spinatus.* The Pliensbachian was named by Oppel in 1858 to replace, and justifiably so, the less appropriate earlier term Liasian introduced by D'Orbigny in 1850.

The name Charmouthian, equally widely used by some authors for exactly the same stratigraphic unit as Pliensbachian, can well be dispensed with as being an unnecessary synonym.

In 1894 Bonarelli proposed the name Domerian for the *Amaltheus margaritatus* and the *Amaltheus spinatus* Zones—the two uppermost Zones of the Pliensbachian Stage. This action led to an ambiguity in the subsequent usage of the term Pliensbachian. Some authors, having accepted Bonarelli's Domerian, have retained the term Pliensbachian for the remaining lower part of the original Pliensbachian of Oppel, whereas others continued to apply the name Pliensbachian to the entire Middle Lias, as used in this report. To eliminate this confusion Lang in 1913 introduced the name Carixian for the lower part of Oppel's Pliensbachian to include the Zones of *Uptonia jamesoni, Tragophylloceras ibex,* and *Prodactylioceras davoei.*

It is suggested here that the name Pliensbachian be retained in its original scope as outlined by Oppel and that the terms Carixian and Domerian be employed as substages for its lower and upper parts respectively.

Toarcian Stage.—This stage represents the Upper Lias and begins at the bottom with the Zone of *Dactylioceras tenuicostatum* and ends above with the Zone of *Lytoceras jurense.* In this scope the Toarcian Stage has been more or less consistently employed since 1850 when it was originally proposed by D'Orbigny.

In 1910 Buckman subdivided the Toarcian Stage into the Whitbian below and the Yeovilian above. Unfortunately, however, according to Buckman's original definition the Yeovilian does not include the *Grammoceras aalense* Zone (or Subzone) which the majority of stratigraphers regard as representing the uppermost Zone of the Upper Lias. It is therefore recommended here that if theYeovilian Substage is to be used at all for the upper part of the Upper Lias it should be revised in conformity with the prevailing usage of Lias to include at the top the *Grammoceras aalense* Subzone of the *Lytoceras jurense* Zone.

DOGGER OR MIDDLE JURASSIC SERIES

General statement.—The Middle Jurassic Series or Dogger, as originally defined by Oppel, begins with the *Leioceras opalinum* Zone below and ends above with the Zone of *Oppelia aspidoides.* It thus includes the Bajocian and Bathonian stages described below.

As can be seen from Table 2 the upper boundary of the Middle Jurassic or Dogger is differently understood by the different stratigraphers. Buch in 1839 was the first to subdivide the Jurassic System into the Lower Jura or Lias, the Middle Jura, and the Upper Jura. Essentially the same subdivisions of the Jurassic were a few years later described by Quenstedt who designated the Lower Jurassic as the Black Jura or Lias, the Middle Jurassic—the Brown Jura, and the Upper Jurassic—the White Jura.

In 1858, Oppel in his classical work on the Jurassic System of England, France, and southwestern Germany, introduced a comprehensive classification of the Jurassic rocks in which the lowest division, the Lias or Lower Jura, was defined in exactly the same scope as the Lias of Buch and Quenstedt, but Oppel's upper two divisions, the Dogger or Middle Jura and the Malm or the Upper Jura, as shown in Table 2, did not correspond to the Middle and Upper Jura respectively of Buch and Quenstedt.

Practically all the British, most of the French, and some of the German geologists have accepted Oppel's classification and have more or less closely adhered to it to the present day. Many German stratigraphers, on the other hand, appear to be following the classification established by Quenstedt. Still others straddle the issue and use Oppel's name Dogger in the scope of the Brown Jura of Quenstedt.

If priority is to be accepted as the deciding factor then the term Dogger of Oppel should be either dropped or its scope should be emended to correspond to that of the Brown Jura of Quenstedt which is the same as the Middle Jura of Buch. Opinions, however, differ as to whether priority should be followed so strictly in stratigraphic nomenclature. Following the present widest usage, especially among the British and French geologists, it is recommended here that the term Dogger, or the Middle Jurassic, be retained as defined by Oppel in 1858. According to this arrangement the Callovian Stage is placed at the bottom of the Upper Jurassic Series or Malm, and the boundary between the Middle and the Upper Jurassic is drawn between the *Oppelia aspidoides* Zone below and the *Macrocephalites macrocephalus* Zone above.

Bajocian Stage.—This stage, as originally defined by D'Orbigny and as generally used today, begins at the bottom with the Zone of *Leioceras opalinum* and ends above with the beds that are placed in the Zone of *Parkinsonia parkinsoni.*

As has been previously stated the introduction of the name Aalenian for the lower part of the original Bajocian caused some confusion and ambiguity. Some workers today use the name Bajocian in the original broad sense, whereas others, particularly in Germany, retain the name Bajocian for the remaining part of the original Bajocian above the Aalenian. It is hoped that interested readers will express their preference on this issue, *i.e.*, whether the name Bajocian should be retained in the original sense or be restricted to the post Aalenian and pre-Bathonian stratigraphic interval.

Bathonian Stage.—This stage includes the Zones of *Oppelia fusca* and of *Oppelia aspidoides.* In this scope the term Bathonian is used today by practically all stratigraphers. The two subdivisions of the Bathonian Stage—the Vesulian Substage for the Zone of *Oppelia fusca* and the Bradfordian Substage for the Zone of *Oppelia aspidoides*—are not listed in the recommended Table, at least for the present. There appears to be little if any need for these terms, each one of which applies to only a single zone.

MALM OR UPPER JURASSIC SERIES

General statement.—As has already been indicated in the discussion of the Middle Jurassic Series or Dogger the Malm or the Upper Jurassic Series begins with the Zone of *Macrocephalites macrocephalus* at the base of the Callovian Stage and ends above with the *Berriasella privasensis* Zone which underlies the lowermost Cretaceous *Spiticeras callistoides* Zone.

There still appears to be some difference of opinion among the workers as to whether an exact correlation of the Upper Jurassic Zones can be established between northwestern Europe, southeastern Europe, and the Russian "boreal" province. These differences of opinion present a stumbling block in the selection of a single standard sequence of zones. Pending further research a composite table of zonal sequences in these different European provinces is offered as Table 1.

Callovian Stage.—This stage was introduced by D'Orbigny in 1850 and in essentially the same scope was designated by Oppel in 1858 as the Kelloway. It begins at the base with the *Macrocephalites macrocephalus* Zone and ends above with the *Peltoceras athleta* Zone. As can be seen from Table 2 there appears to be some disagreement among the workers regarding the scope of the Callovian Stage. The most widely accepted usage, however, agrees with the original definition of D'Orbigny and the subsequent redefinition by Oppel. In this scope it is recommended in the present report.

A particular detail on which there appears to be some disagreement among the more recent workers is the position of the *Quenstedioceras lamberti* Zone. Some authors place the *lamberti* Zone at the top of the Callovian, whereas others assign it to the next higher Oxfordian Stage. There are also those who leave it out entirely from the zonal sequence of the Upper Jurassic. In accordance with the most widely accepted usage the *Quenstedioceras lamberti* Zone is here placed at the base of the Oxfordian Stage.

Oxfordian Stage.—The wide diversity of opinion as to the scope of the Oxfordian Stage is strikingly shown in Table 2. The usage recommended in the present report is essentially that which was established by Oppel in 1858 and accepted by Waagen in 1865. The Oxfordian Stage begins below with the *Quenstedioceras lamberti* Zone and ends above with the Zone of *Peltoceras bimammatum.*

Kimmeridgian Stage.—As has been pointed out in the earlier part of this article there has been some confusion with regard to the use of the name Kimmeridgian as a Stage term. The conflicting use of this term has been traced to the failure on the part of some workers to discriminate the Stage terms (or units) from the formational terms (or units). Further confusion in the use of the Kimmeridgian Stage is due to the unfortunate introduction by D'Orbigny of the Corallian Stage above the Kimmeridgian. The Corallian Stage, named primarily for a coralline facies, has been eliminated from the list of the recommended Stages, all of which are based on the ammonitic facies.

As recommended in this report the Kimmeridgian Stage begins below with the *Sutneria platynota* Zone and ends at the top with the Zone of *Aulacostephanus eudoxus*.

Tithonian Stage.—The Tithonian Stage was established by Oppel in 1865 for the beds intermediate between the Kimmeridgian and the lowest Cretaceous. It is clear from Oppel's original definition that the Tithonian Stage should include the so-called Portlandian beds and the marine beds equivalent of Purbeck of England. From the large number of species which Oppel listed as characteristic of the Tithonian Stage it is clear that the boundary between the Kimmeridgian and Tithonian stages was placed by him between the *Aulacostephanus pseudomutabilis* Zone below and the *Gravesia gravesiana* Zone above. In this scope the Tithonian Stage has been widely used by the majority of authors and in this scope it is recommended in the present report.

Spath's restriction of the Tithonian Stage in 1923 to the beds above the *Titanites giganteus* Zone and below the Cretaceous just because Oppel "clearly meant it to be used for strata that are transitional to, or indicate the dawn of, Cretaceous," cannot be accepted. It can hardly be called a sound procedure to use such a subjective criterion for a revision of the Stage. It is not difficult to imagine someone a few years hence claiming that only a small part of Spath's "transitional" beds are really transitional and advocating further restriction of the scope of the Tithonian Stage. It is difficult to see how under such circumstances any stability of nomenclature can be achieved.

Furthermore, the beds above the *Titanites giganteus* Zone and below the Cretaceous have already been designated by Pavlov in 1896 as the Aquilonian Stage (or Substage). Spath's revision of the Tithonian, therefore, may be rejected simply because there is no justifiable need for another term for exactly the same part of the section for which Pavlov had proposed the name Aquilonian.

The terms Bononian and Aquilonian, introduced by Pavlov in 1896, are proposed in this report as Substages of the Tithonian Stage.

For the lower Substage the name Bononian is preferred to Portlandian—an older but much more ambiguous term. Opinions among the workers as to the stratigraphic scope of the Portlandian are so widely divergent that the term has become meaningless. For this reason it is better to supplant it by the term Bononian which was precisely defined at the time of its introduction and which has been used little if at all in any other sense. (*See* Table 2.)

For the upper substage the term Aquilonian is preferred to the Purbeckian. The Purbeckian section is composed of brackish-water and

continental deposits and therefore provides no means of correlation with the ammonite-bearing beds.

Only a selected list of a large number of references consulted in the course of the present investigation is appended. An exhaustive list of papers dealing with the Jurassic System may be found in recent publications by W. J. Arkell, L. F. Spath, Frédéric Roman, and others.

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Stanford University, California.

Manuscript received by the Secretary of the Society, January 29, 1941.

RECOMMENDED STANDARD OF THE JURASSIC S

SERIES	STAGES and SUBSTAGES		ZONES and SUBZONES	
MALM or UPPER JURASSIC	TITHONIAN	AQUILONIAN	Berriasella privasensis	Berriasella delphinen Protacanthodiscus ch
		BONONIAN	Titanites giganteus	Virgatosphinctes tenui Proniceras idoceroide
			Provirgatites scythicus	
			Subplanites contignus	Pavlovia rotunda Virgatosphinctes pseu
			Neochetoceras steraspis	Subplanites viminous Waagenia hybonota Waagenia beckeri
	KIMMERIDGIAN		Aulacostephanus eudoxus	Sutneria subeumela Aulacostephanus pseudo
			Streblites tenuilobatus	Idoceras balderum Ataxioceras polyplocum
			Sutneria platynota	Sutneria platynota Idoceras planulum
	OXFORDIAN		Peltoceras bimammatum	
			Aspidoceras hypselum	
			Peltoceras transversarium	
			Aspidoceras perarmatum	
			Cardioceras cordatum	
			Quenstedioceras mariae	
			Quenstedioceras lamberti	
	CALLOVIAN		Peltoceras athleta or Kosmoceras ornatum	
			Reineckeia anceps	Kosmoceras castor Kosmoceras jason
			Sigaloceras calloviensis	
			Macrocephalites macrocephalus	
DOGGER or MIDDLE JURASSIC	BATHONIAN		Oppelia aspidoides	Kamptokephalites harvey Clydoniceras discus Wagnericeras wagneri
			Oppelia fusca	Morrisites morrisi Siemiradzkia aurigera
	BAJOCIAN		Parkinsonia parkinsoni	Zigzagiceras zigzag Parkinsonia schloenbachi Strigoceras truellei Garantiana garantiana
			Strenoceras subfurcatum	
			Stephanoceras humphriesianum	Teloceras blagdeni Stephanoceras humphriesi Witchellia romani
			Otoites sauzei	
			Sonninia sowerbyi	Witchellia laeviuscula Shirburnia trigonalis Hyperlioceras discites
			Ludwigia murchisonae	Ludwigella concava Brasilia bradfordensis Ludwigia murchisonae Staufenia staufensis Ludwigia discoidea Ludwigia sehndensis
			Leioceras opalinum	Ludwigia toluteria Leioceras sinon Leioceras costosum Leioceras opalinum Lytoceras torulosum
UPPER LIAS	TOARCIAN		Lytoceras jurense	Grammoceras (Pleydellia) Dumortieria radiosa Phlyseogrammoceras dispa Grammoceras fallaciosum Grammoceras striatulum Haugia variabilis
			Hildoceras bifrons	Dactylioceras braunianum Coeloceras (Paracoeloceras) Frechiella subcarinata
			Harpoceras falciferum	Harpoceras falciferum Harpoceras exaratum
			Dactylioceras tenuicostatum	Dactylioceras athleticum Dactylioceras tenuicosta Tiltoniceras acutum
LIAS		DOMERIAN	Paltopleuroceras spinatum	Paltopleuroceras spinatu Paltopleuroceras hochste
			Amaltheus margaritatus	Amaltheus margaritatus Amaltheus nodifer
	PLIENSBACHIAN			Cisticoceras figulinum Becheiceras bechei

SYSTEM

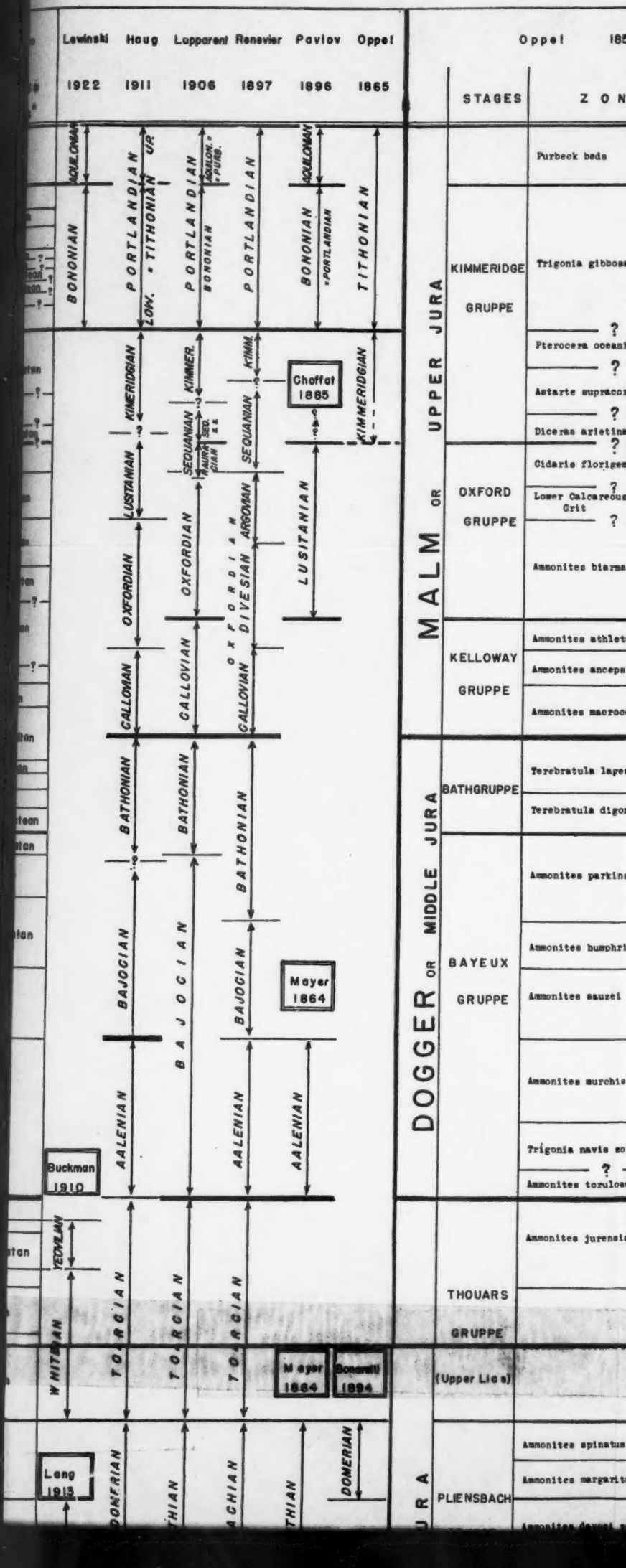
STAGES AS

BZONES

Beringer 1937	Arkell 1933 etc.	Spath 1933 etc.	G. M. Davies 1932	A. Morley Davies 1929	Buckman 1927-1930 "Ages"	Lewinski 1922	Haug 1911	Lapparent 1906
PURBECK.	PURBECK. TITHON.	TITHON.	PURBECK. TITHON.	PURBECK. AQUILON.		AQUILONIAN		
PORTLANDIAN	PORTLAND.	PORTLAND.	PORTLAND.	PORTLAND.	Gigantitan	BONONIAN	PORTLANDIAN	PORTLANDIAN
					Behemothan			
					Virgatitan			
					Paravirgatitan ?			
	"KIMERIDGE CLAY"				Aladrgatitan ?			
					Holcosphinctean ?			
					Aulacosphinctean ?			
					Mazapillitan ?			
				BONONIAN	Gravesian			
KIMMER.		KIMMERIDGIAN	KIMMERIDGIAN	KIM.	Physodoceratan		KIMERIDGIAN	KIMMER.
			SEQUANIAN HAVRIAN	SEQUANIAN				
LUSIT.	SEQUAN.				Rasenian ?			
					Prionodoceratan ?			SEQUANIAN
				RAUR.	Ringsteadian		LUSITANIAN	
OXFORDIAN	ARGO. RAUR.	OXFORDIAN	OXFORDIAN	ARGOVIAN	Periphinctean			
	CORALLIAN	"ARGOVIAN" "CORALLIAN"	ARGOVIAN		Cardioceratan		OXFORDIAN	OXFORDIAN
				DIVESIAN	Vértumniceratan			
		DIVESIAN	DIVESIAN		Kosmoceratan ?			
CALLOVIAN	CALLOVIAN	CALLOVIAN	CALL. ?	CALLOVIAN	Reineckeian ?		CALLOVIAN	CALLOVIAN
					Proplanulitan			
					Macrocephalitan			
BATHONIAN	BATHONIAN	BATHONIAN	BATHONIAN	BATHONIAN	Clydoniceratan		BATHONIAN	BATHONIAN
					Oxyceritan			
					Tulitan			
					Gracilisphinctean			
	VESULIAN ?		VESULIAN	VESULIAN ?	Zigzagiceratan			
BAJOCIAN		BAJOCIAN			Parkinsonian		BAJOCIAN	BAJOCIAN
	BAJOCIAN		BAJOCIAN AALENIAN	BAJOCIAN	Stepheoceratan			
					Sonninian			
AALENIAN	AALENIAN			AALENIAN	Ludwigian	Buckman 1910	AALENIAN	
	YEOVIL ?			YEOVIL ?	Conavarian			
					Dumortierian	YEOVILIAN		
					Grammoceratan			
TOARCIAN	WHITBIAN	TOARCIAN	TOARCIAN	WHITBIAN	Haugian	WHITBIAN	TOARCIAN	TOARCIAN
					Hildoceratan			
					Harpoceratan			
DOMERIAN	DOMERIAN "UP. PLIENSB."	DOMERIAN	DOMERIAN	DOMERIAN	Amaltheian	Lang 1913	DOMERIAN	
					Liparoceratan			

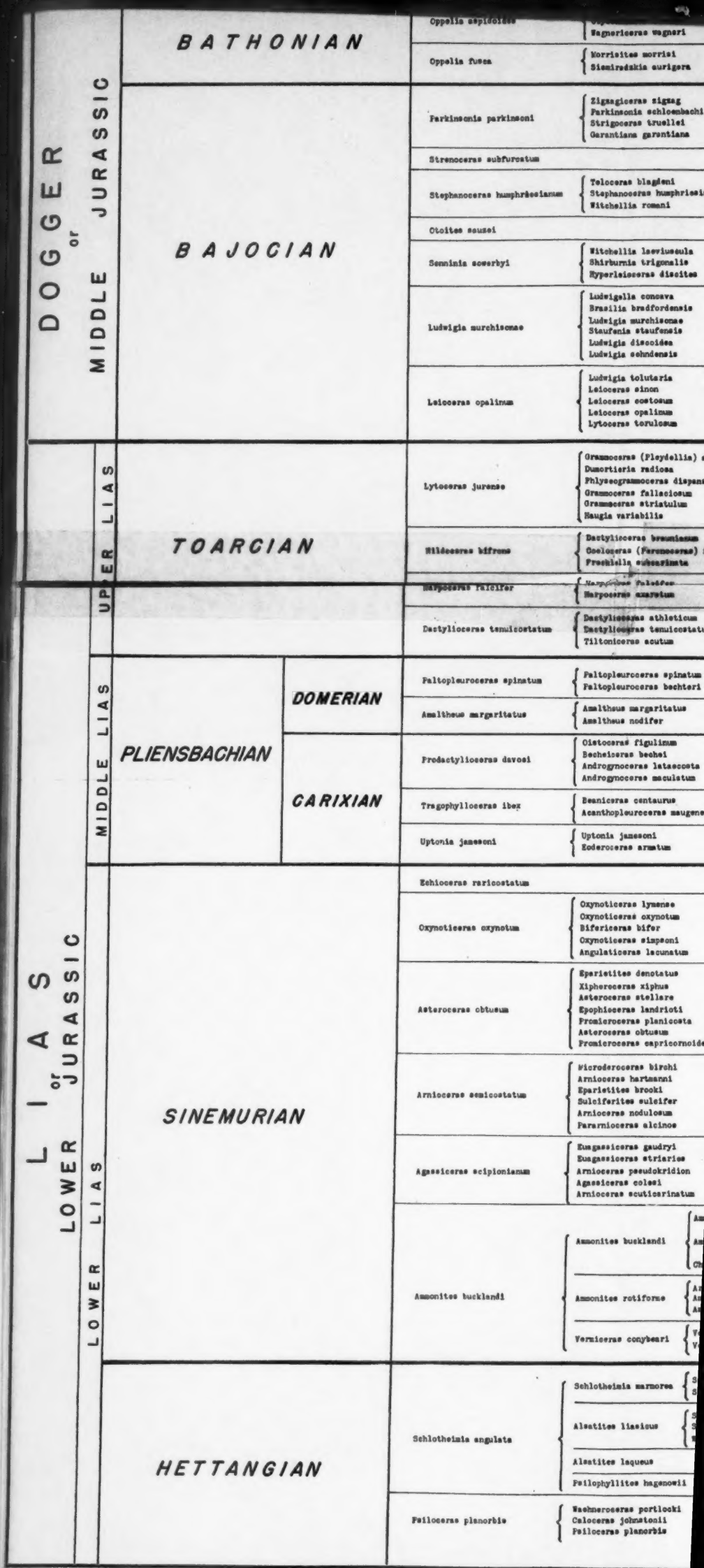
lla delphinensis
thodiscus chaperi
phinctes tenuistriata
as idoceroides
rotunda
phinctes pseudoulmensis
tes vimineus
hybonota
beckeri
subumela
ephanus pseudomutabilis
baldorum
ras polyplocum
platynota
planulum
as castor
as jason
halites herveyi
eras discus
eras wagneri
es morrisi
kia aurigera
eras zigzag
nia schloenbachi
ras truellei
ra garantiana
e blagdeni
ceras humphriesianum
ia romani
ia laeviuscula
ia trigonalis
ceras discites
la concava
bradfordensis
murchisonae
staufensis
discoides
schndensis
toluteria
sinon
costosum
opalinum
torulosum
as (Pleydellia) aalense
ia radiosa
mmoceras dispansum
as fallaciosum
as striatulum
riabilis
eras braunianum
s (Paroncoceras) fibulatum
subcarinata
e falciferum
e exaratum
eras athleticum
eras tenuicostatum
eras acutum
roceras spinatum
roceras bechteri
margaritatus
nodifer
figulinum
as bechei

STAGES AS DEFINED BY DIFFERENT

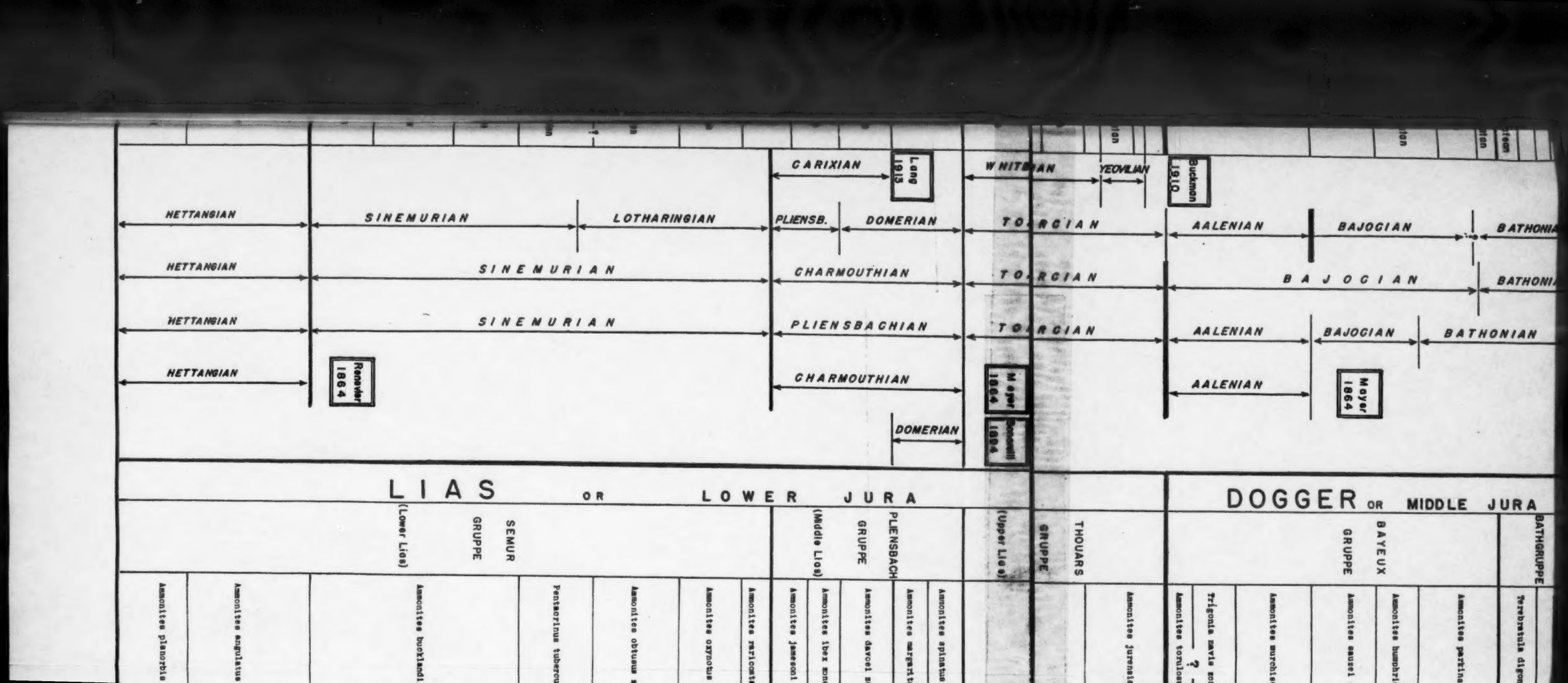


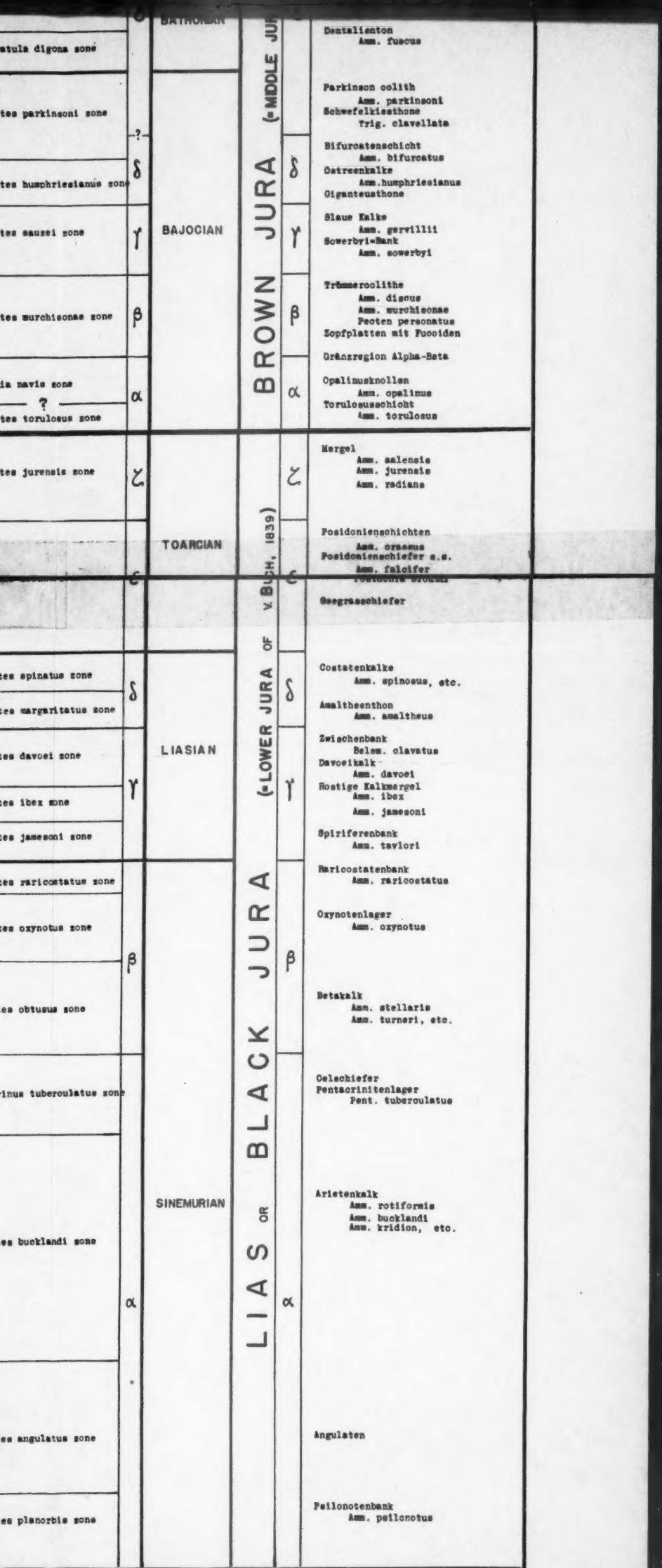
NT AUTHORS

1855 - 1858 ZONES	Q's sym.	D'Orbigny 1842 etc.	Quenstedt 1843 etc.		
eck beds					
onia gibbosa zone		PORTLANDIAN	WHITE JURA (= UPPER JURA OF v. BUCH)	?	
?	?			ζ	Kalkplatten mit Krebsscheren
					Lithographische Schiefer
?				?	? Mergel ?
				ε	Massenkalk und Dolomit
pocera oceani zone	ζ	KIMMERIDGIAN		δ	Geschichtete Felsenkalke
?					Amm. flexuosus
rte supracorallina zone	?			γ	Planulaten - Amm. dentatus
?	ε	?			
ras arietina zone	?	CORALLIAN		β	Wohlgeschichtete Kalkbanke, et
?	δ				
ris florigemma zone	?	?			
r Calcareous Grit — Scyphien-kalke	γ β				
?					
	α	OXFORDIAN		α	Impressa Mergel
hites biarmatus zone	?				Amm. alternans
	?	?			
hites athleta zone	ζ		BROWN JURA (= MIDDLE JURA OF v. BUCH)	ζ	Amm. ornatus
hites anceps zone					Amm. jason
hites macrocephalus zone		CALLOVIAN			Macrocephalus oolith
					Amm. macrocephalus
		?			
pratula lagenalis zone	ε	BATHONIAN		ε	Dentalienton
pratula digona zone					Amm. fuscus
hites parkinsoni zone					Parkinson oolith
					Amm. parkinsoni
	?				Schwefelkiesthone
					Trig. clavellata
hites humphriesianus zone	δ			δ	Bifurcatenschicht
					Amm. bifurcatus
					Ostreenkalke
					Amm. humphriesianus
					Gigganteusthone
hites sauzei zone	γ	BAJOCIAN		γ	Blaue Kalke
					Amm. gervillii
					Sowerbyi-Bank
					Amm. sowerbyi
hites murchisonae zone	β			β	Trümmeroolithe
					Amm. discus
					Amm. murchisonae
					Peoten personatus
					Zopfplatten mit Fucoiden
					Grenzregion Alpha-Beta
nia navis zone	α			α	Opalinusknollen
?					Amm. opalinus
hites torulosus zone					Torulosusschicht
					Amm. torulosus
hites jurensis zone	ζ	TOARCIAN	LOWER JURA OF v. BUCH, 1839	ζ	Mergel
					Amm. salensis
					Amm. jurensis
					Amm. radians
					Posidonienschichten
					Amm. crassus
	ε			ε	Posidonienschiefer e.s.
					Amm. falcifer
					Seegrasschiefer
hites spinatus zone	δ	LIASIAN		δ	Costatenkalke
					Amm. spinosus, etc.
hites margaritatus zone					Amaltheenthon
					Amm. amaltheus
					Zwischenbank
					Belem. clavatus


DOGGER or MIDDLE JURASSIC
BATHONIAN
Oppelia aspidoides
Wagnericeras wagneri
Oppelia fusca
Morrisites morrisi
Siemiradzkia aurigera
BAJOCIAN
Parkinsonia parkinsoni
Zigzagiceras zigzag
Parkinsonia schloembachi
Strigoceras truellei
Garantiana garantiana
Strenoceras subfurcatum
Stephanoceras humphriesianum
Teloceras blagdeni
Stephanoceras humphriesi
Witchellia romani
Otoites sauzei
Sonninia sowerbyi
Witchellia laeviuscula
Shirburnia trigonalis
Hyperlioceras discites
Ludwigia murchisonae
Ludwigella concava
Brasilia bradfordensis
Ludwigia murchisonae
Staufenia staufensis
Ludwigia discoides
Ludwigia sehndensis
Leioceras opalinum
Ludwigia tolutaria
Leioceras sinon
Leioceras costosum
Leioceras opalinum
Lytoceras torulosum
UPPER LIAS
TOARCIAN
Lytoceras jurense
Grammoceras (Pleydellia)
Dumortieria radiosa
Phlyseogrammoceras dispans
Grammoceras fallaciosum
Grammoceras striatulum
Haugia variabilis
Hildoceras bifrons
Dactylioceras braunianum
Coeloceras (Paracoeloceras)
Frechiella subcarinata
Harpoceras falciferum
Harpoceras exaratum
Dactylioceras tenuicostatum
Dactylioceras athleticum
Dactylioceras tenuicostatum
Tiltoniceras acutum
MIDDLE LIAS
PLIENSBACHIAN
DOMERIAN
Paltopleuroceras spinatum
Paltopleuroceras spinatum
Paltopleuroceras hechteri
Amaltheus margaritatus
Amaltheus margaritatus
Amaltheus nodifer
CARIXIAN
Prodactylioceras davoei
Oistoceras figulinum
Becheiceras bechei
Androgynoceras lataecosta
Androgynoceras maculatum
Tragophylloceras ibex
Beaniceras centaurus
Acanthopleuroceras maugenesti
Uptonia jamesoni
Uptonia jamesoni
Echioceras raricostatum
LIAS or LOWER JURASSIC
LOWER LIAS
SINEMURIAN
Oxynoticeras oxynotum
Oxynoticeras lymense
Oxynoticeras oxynotum
Bifericeras bifer
Oxynoticeras simpsoni
Angulaticeras lacunatum
Asteroceras obtusum
Eparietites denotatus
Xipheroceras xiphus
Asteroceras stellare
Epophioceras landrioti
Promicroceras planicosta
Asteroceras obtusum
Promicroceras capricornoides
Arnioceras semicostatum
Microderoceras birchi
Arnioceras hartmanni
Eparietites brooki
Sulciferites sulcifer
Arnioceras nodulosum
Paramioceras alcinoe
Agassiceras scipionianum
Euagassiceras gaudryi
Euagassiceras striarius
Arnioceras pseudokridion
Agassiceras colosi
Arnioceras scuticarinatum
Ammonites bucklandi
Ammonites bucklandi
Ammonites rotiformis
Vermiceras conybeari
HETTANGIAN
Schlotheimia angulata
Schlotheimia marmorea
Alsatites liasicus
Alsatites laqueus
Psilophyllites hagenowii
Psiloceras planorbis
Waehneroceras portlocki
Caloceras johnstonii
Psiloceras planorbis

| | | |


HETTANGIAN
SINEMURIAN
LOTHARINGIAN
CARIXIAN
Lang 1913
PLIENSB.
DOMERIAN
WHITBIAN
YEOVILIAN
Buckman 1910
TORCIAN
AALENIAN
BAJOCIAN
BATHONIAN
HETTANGIAN
SINEMURIAN
CHARMOUTHIAN
TORCIAN
BAJOCIAN
BATHONIAN
HETTANGIAN
SINEMURIAN
PLIENSBACHIAN
TORCIAN
AALENIAN
BAJOCIAN
BATHONIAN
HETTANGIAN
Renevier 1864
CHARMOUTHIAN
Meyer 1864
Bonarelli 1894
AALENIAN
Meyer 1864
DOMERIAN
LIAS OR LOWER JURA
DOGGER OR MIDDLE JURA
SEMUR GRUPPE
(Lower Lias)
PLIENSBACH GRUPPE
(Middle Lias)
(Upper Lias)
THOUARS GRUPPE
BAYEUX GRUPPE
BATHGRUPPE
Ammonites planorbis
Ammonites angulatus
Ammonites bucklandi
Pentacrinus tuberculatus
Ammonites obtusus
Ammonites oxynotus
Ammonites raricostatus
Ammonites jamesoni
Ammonites ibex
Ammonites davoei
Ammonites margaritatus
Ammonites spinatus
Ammonites jurensis
Ammonites torulosus
Trigonia navis
Ammonites murchisonae
Ammonites sauzei
Ammonites humphriesianus
Ammonites parkinsoni
Terebratula digona


BATHONIAN
atula digona zone
tes parkinsoni zone
tes humphriesianus zone
tes sauzei zone
tes murchisonae zone
ia navis zone
?
tes torulosus zone
BAJOCIAN
BROWN JURA (= MIDDLE JU
Dentalienton
Amm. fuscus
Parkinson oolith
Amm. parkinsoni
Schwefelkiesthone
Trig. clavellata
Bifurcatenschicht
Amm. bifurcatus
Ostreenkalke
Amm. humphriesianus
Giganteusthone
Blaue Kalke
Amm. gervillii
Sowerbyi-Bank
Amm. sowerbyi
Trümmeroolithe
Amm. discus
Amm. murchisonae
Pecten personatus
Zopfplatten mit Fucoiden
Grenzregion Alpha-Beta
Opalinusknollen
Amm. opalinus
Torulosusschicht
Amm. torulosus
tes jurensis zone
TOARCIAN
BROWN JURA (= LOWER JURA OF v. BUCH, 1839)
Mergel
Amm. aalensis
Amm. jurensis
Amm. radians
Posidonienschichten
Amm. crassus
Posidonienschiefer s.s.
Amm. falcifer
Seegrasschiefer
tes spinatus zone
tes margaritatus zone
tes davoei zone
tes ibex zone
tes jamesoni zone
LIASIAN
Costatenkalke
Amm. spinosus, etc.
Amaltheenthon
Amm. amaltheus
Zwischenbank
Belem. clavatus
Davoeikalk
Amm. davoei
Rostige Kalkmergel
Amm. ibex
Amm. jamesoni
Spiriferenbank
Amm. tavlori
tes raricostatus zone
tes oxynotus zone
tes obtusus zone
Raricostatenbank
Amm. raricostatus
Oxynotenlager
Amm. oxynotus
Betakalk
Amm. stellaris
Amm. turneri, etc.
inus tuberculatus zone
es bucklandi zone
es angulatus zone
es planorbis zone
SINEMURIAN
LIAS OR BLACK JURA
Oelschiefer
Pentacrinitenlager
Pent. tuberculatus
Arietenkalk
Amm. rotiformis
Amm. bucklandi
Amm. kridion, etc.
Angulaten
Psilonotenbank
Amm. psilonotus

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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STRUCTURE OF THE ELK MOUNTAIN DISTRICT, CARBON COUNTY, WYOMING

BY R. H. BECKWITH

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2. Tectonic map of the region around the north end of the Medicine Bow Range . 1482

ABSTRACT

The Elk Mountain district is at the northwest end of the Medicine Bow Range, where trend changes from northerly in the Southern Rockies to westerly in the Central Rockies of Wyoming. Deformation of pre-Cambrian crystallines and 17,000 feet of predominantly shaly Paleozoic and Mesozoic beds proceeded during and after the deposition of many thousand feet of Upper Cretaceous and Fort Union beds in the basin to the north. Folds trend generally north, and their axial planes dip west. The following unusual features are described and discussed in terms of genesis:

(1) Two westward-dipping thrusts pass laterally northward into west-striking tears, which cut across folds; deformation in a block on one side of a tear was more or less independent of that in the other block. (2) A thrust cuts diagonally across a major syncline and anticline. (3) Marked rotatory movement is shown by a northeast-striking tear, which terminates upward at a thrust, and by several smaller tears cutting across slices between thrusts. (4) A fault having the characteristics of a subsequent shear thrust is probably a combined break and erosion thrust. (5) A bedding thrust in the bentonite-bearing Mowry shale of the gentle west flank of an anticline crosses the axis and offsets the axial plane.

Oil and gas possibilities of the district and the possible effects of bedding thrusts on oil traps are discussed.

The structural pattern of the district and surrounding region (Pl. 2) diverges greatly from that produced by uniform stress applied to uniform rocks. This is attributed to effective compression acting in different directions and to the influence of such features as bedding, schistosity, contacts of igneous bodies, and faults in the pre-Cambrian.

INTRODUCTION AND ACKNOWLEDGMENTS

Three anticlinal mountain ranges extend into southeastern Wyoming (Fig. 1) from the south—the Laramie, Medicine Bow, and the Sierra Madre. The Sweetwater Arch, a broad belt of pre-Cambrian rocks, extends west-northwest for 100 miles from a point 35 miles north of the Medicine Bow Range and joins the Wind River Range in west-central Wyoming.

The Elk Mountain district at the northwest end of the Medicine Bow Range occupies a position of abrupt change from the northerly structural trends of the Southern Rockies to the westerly trends of central Wyoming. The present study was undertaken by the Geological Survey of Wyoming

to obtain detailed information on this change and to extend present knowledge of relations of folds, thrusts, and tear faults. It was also hoped that certain structural relations shown in beds of the lower part

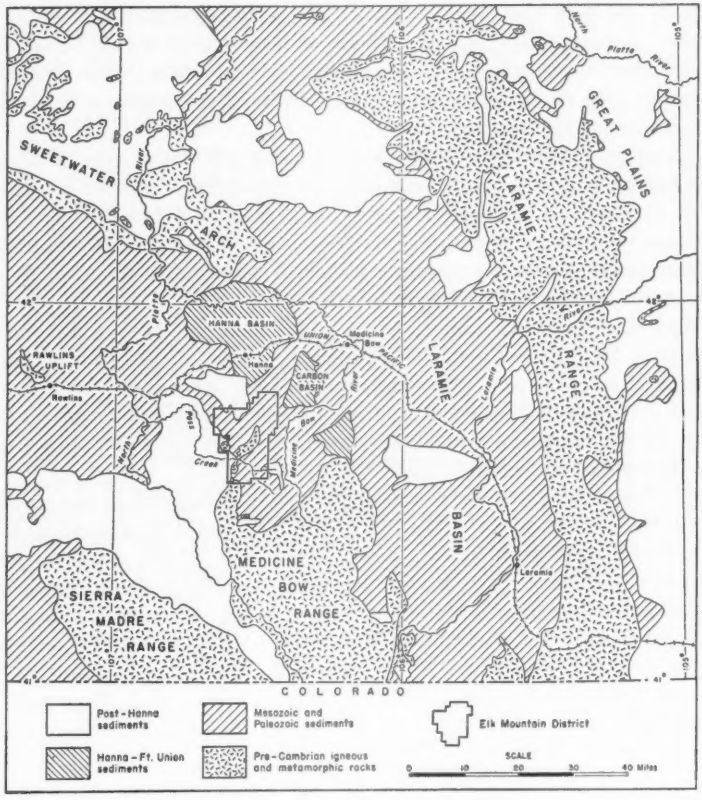

FIGURE 1.—*Index map of southeastern Wyoming*

of the stratigraphic column in areas of good exposures would facilitate interpretation in other areas where only isolated surface observations or subsurface data can be obtained, as in some of the oil fields in the surrounding region.

The area covered by the writer was first mapped in very generalized form by the Fortieth Parallel Survey (King, 1876, Map I). Veatch (1907) investigated the coal-bearing rocks in the northern part but

grouped as a unit all of the older sediments not containing commercial coal. The map by Dobbin and associates (1929a) on a scale of 1 inch to a mile includes the Hanna and Carbon basins and adjacent territory. A strip one mile wide at the north end of the writer's map (Pl. 1) overlaps their area. Neely's reconnaissance of the north end of the Medicine Bow Range (Neely, 1934) included Coad Mountain and the southeast end of Elk Mountain. Isberg (1937) made a reconnaisance map of the west and north sides of Elk Mountain and a considerable area to the west.

All of the Elk Mountain district, except a strip 2 to 4 miles wide along the eastern border, is included in the Fort Steele sheet, a reconnaissance map by the U. S. Geological Survey on a scale of 1:125,000 and with contour intervals of 25 and 50 feet. Topography is not tied in with land subdivisions and is not accurately enough represented to serve as a base map for detailed geology.

Approximately 135 square miles is included in the mapped area (Pl. 1). With the exception of the interior parts of the pre-Cambrian cores of the mountains, the mapping was done with plane table on a scale of 1 inch=1000 feet for the northwestern third and 1 inch=1500 feet for the remainder. Field work was started in the summer of 1938 and completed in the fall of 1940.

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TOPOGRAPHY AND DRAINAGE

The most prominent topographic feature of the district (Pl. 1) is the high broken ridge formed by the crystalline core of the western branch of the Medicine Bow Range. It extends from the southwest corner to the north-central part of the district. Most of the ridge crest is at elevations above 9000 feet and attains a maximum of 11,162 feet at the peak of Elk Mountain. The ridge is broken by two gaps. The elevation of the southern one, Oberg Pass between Pennock Mountain and Coad Mountain, is 8200 feet. Pass Creek has cut a canyon between Coad Mountain and Elk Mountain forming a second gap at an elevation of

7300 feet. All three mountains are markedly asymmetric, having steeper eastern faces.

From the eastern bases of the mountains alluvium-covered pediments slope eastward at angles as high as 7° and gradually flatten toward the Pass Creek Basin and the valley of the Medicine Bow River. At the north base of Elk Mountain the Cloverly conglomerate forms a prominent ridge that swings around to the west and south. The sandstones of the Mesaverde formation form another ridge up to 500 feet high extending northeastward across the northern part of the district. From the bases of both ridges a gently rolling alluvium-covered surface slopes toward the alkali lakes in the lower part of the basin of Halleck Creek at elevations around 7300 feet.

West of Elk Mountain the Cloverly conglomerate forms two high ridges on the flanks of a northward-trending syncline. Strike valleys on the adjoining shale outcrops drain north and south into Rattlesnake Creek and Pass Creek.

Sheephead Mountain in the northwestern part of the district is a strongly asymmetric mountain rising to 8700 feet. The east side is very steep and, in some places, is nearly vertical for several hundred feet. The slope toward the north is more gradual and is broken by a ridge of Cloverly conglomerate. About a mile north Rattlesnake Creek flows in a bend roughly parallel to the Cloverly ridge. Northeast and northwest from the bend the surface rises abruptly about 500 feet to the crest of the ridge formed by the Mesaverde sandstone. Sheephead Mountain is sharply cut off on the south by the valley of Hat Creek, which narrows to a canyon several hundred feet deep in crossing the upturned resistant sediments at the southeast end of the mountain.

Bear Butte is composed of granites and gneisses. The eastern part near the watershed is carved into sharp crags and pinnacles. Farther west the topography is more subdued and merges into the gently rolling surface cut on the non-resistant Tertiary beds, which extend westward from the mountains to the valley of the North Platte.

The northeastern part of the district is drained by a few small permanent streams and a number of intermittent streams tributary to the Medicine Bow River (Fig. 1). The northwestern part is drained by Rattlesnake Creek and Hat Creek, which enter Pass Creek a few miles west of the mountains. A large part of the embayment in the north end of the Medicine Bow Range drains into Pass Creek, which flows through the canyon between Elk Mountain and Coad Mountain and enters the North Platte about 20 miles to the northwest. That a continuous cover of younger Tertiary sediments once extended through the gap is indicated by the large areas of Tertiary sediments in the upper Pass

Creek Basin, the small remnants north and south of the east end of the canyon (Pl. 1) occurring as high as 1000 feet above the present flood plain, and the extensive area west of the mountains. Pass Creek is probably a superimposed stream. No Tertiary sediments were found above 9000 feet, and it is therefore improbable that the mountains were completely covered. It is suggested that before the deposition of the younger Tertiary erosion had cut a saddle along the Bear Butte fault between Elk Mountain and Coad Mountain.

Glaciation has played only a very minor part in land sculpture. A few of the stream valleys heading in the higher parts of Elk Mountain have broad cirquelike heads, and some of the material mapped as alluvium has the characteristics of glacial material. The V-shaped profiles of valleys below 9500 feet indicate that ice did not penetrate far below this elevation.

PRE-CAMBRIAN ROCKS

Detailed studies of the pre-Cambrian rocks were not made. In working around the borders of pre-Cambrian areas, however, various features were observed. The alluvial material along short streams also provided information on the rocks along their courses. The writer believes that structures in the pre-Cambrian complex exerted a considerable influence on those formed during the Laramide orogeny, and the following observations are therefore given.

The core of the Medicine Bow Range, 10 to 15 miles southeast of the embayment in the north end, contains an extensive area of metamorphic rocks. Blackwelder (1926) describes a succession of quartzites, dolomitic marbles, slates, schists, and metamorphosed pyroclastics occupying an area 5 miles wide and extending northeast for at least 12 miles. The succession is about 24,000 feet thick. The metamorphosed sediments and pyroclastics are cut by metadiabase and granite dikes and are enclosed in a complex of gneisses, schists, and granites. Beds strike northeast and dip steeply to the southeast. A major strike fault apparently has a high dip. A number of nearly vertical dip faults offset beds hundreds of feet.

The oldest rocks in the Elk Mountain district are hornblende schists and metadiabases; they were probably originally basalt flows and diabase sills and dikes. Some of the more micaceous schists may have been argillaceous sedimentary rocks. The writer has seen no evidence of metamorphosed sediments such as quartzites and marbles in the exposed parts of the pre-Cambrian, and the gulches heading in the mountains of the district do not carry quartzite pebbles. The abundant quartzite boulders and pebbles in the Tertiary sediments and alluvium of the Pass Creek Basin

are undoubtedly derived from the known areas of quartzites to the southeast. The schists and metadiabases have been invaded by granites, predominantly of the potash variety, and have been stoped, absorbed, and injected to form gneisses. In addition there are extensive areas of massive granites.

STRATIGRAPHY

GENERAL STATEMENT

The stratigraphic section (Table 1) includes approximately 15,000 feet of Cambrian to Upper Cretaceous beds. Disconformities cut out part of a system or several systems, but in all cases the tilting and accompanying erosion were insufficient to produce noticeable angular discordance of beds. There are two angular unconformities, one at the base of the Madison or at the base of the thin underlying Cambrian quartzite lenses, and the other at the base of the North Park, which laps onto all older rocks.

Thicknesses of the Madison, Fountain, and Tensleep were obtained from widths of outcrop in the SE ¼ sec. 4, T. 19 N., R. 82 W., where the beds are nearly vertical. The Chugwater was measured by steel tape and clinometer in the SW ¼ sec. 28 and SE ¼ sec. 26, T. 20 N., R. 82 W. A closely comparable thickness was obtained from mapped width of outcrop and dips in the southern part of the district in the NW ¼ sec. 11, T. 18 N., R. 82 W. Thicknesses of the formations from Jelm to Carlile inclusive were measured by steel tape and clinometer in the W ½ sec. 29, T. 20 N., R. 82 W; those of the Niobrara to Lewis inclusive were obtained by scaling widths of mapped outcrops and dips in the northern part of the district at places where the two boundaries of the formation crop out at nearly the same elevation. The figures for the Medicine Bow and North Park are those of Dobbin and associates (1929a, p. 9-10) in the adjacent area to the north.

NOTES ON NOMENCLATURE

The stratigraphy is well known. The following notes are given to correlate stratigraphic terminology with that used in surrounding territory.

The quartzite lenses lying on the pre-Cambrian are erosional remnants left by pre-Madison erosion of Middle Cambrian quartzites, which attain a thickness of 412 feet at Rawlins 35 miles to the northwest (Dobbin, *et al.*, 1928, p. 176).

The Madison limestone, 298 feet thick in the northern part of the Rawlins uplift (Dobbin, *et al.*, 1928, p. 176), thins to the southeast and is absent 15 miles south and east of Elk Mountain (Neely, 1934, p. 4).

The limestones and redbeds here called Fountain lie between the Madison and Tensleep in the same position as beds of comparable lithology and thickness called

TABLE 1.—*Geologic formations in the Elk Mountain district*

Age	Formation	Thickness (feet)	Characteristics
Quaternary (?)		0–50 (?)	Floodplain and pediment gravels, land-slip debris, and probably some thin remnants of Tertiary sediments.
Miocene	North Park formation	0–400 (?)	Yellowish partially consolidated arkosic gravel and sand, and gray clay. Several hundred feet above the base are white unconsolidated volcanic ashes and silicified ash beds resembling limestones. Continental.
	Unconformity		
Upper Cretaceous	Medicine Bow formation	4000–6200	Gray to brown sandstones, light-colored to gray and carbonaceous shales, and thin coals.
	Lewis shale	3000	Predominantly dark-gray shale. Several bands of light-gray sandstone and sandy shale contain hard brown "ironstone" concretions up to 5 feet across. Marine.
	Mesaverde formation	3500	Brown, gray, and white sandstones alternating with light-gray to carbonaceous shales. Thin coals in the upper part. Lower part marine. Forms a ridge.
	Steele shale	2400	Predominantly dark-gray shale. In the upper 600 to 700 feet are the dark-gray thin sandstones of the Shannon member. Marine.
	Niobrara formation	700	Buff shaly limestones at the base, near the middle, and at the top. Above and below the middle limestone are gray to black shales. Marine.
	Carlile shale	420	Dark-gray to black shales containing light-gray calcareous concretions up to 2 feet across. Marine.
	Frontier formation	650	Black shale. At the top is the brown Wall Creek sandstone member about 30 feet thick. In the lower part are black septarian concretions up to 2 feet across. Marine.
	Mowry shale	165	Black siliceous shale that weathers silver gray; contains bentonite seams up to 2 feet thick. Marine.
	Thermopolis shale	110	Dark-gray to black shales with thin sandy beds. The "Muddy" sandstone, a few feet thick, is locally present at the top. Marine.
Lower Cretaceous	Cloverly formation	130	Upper member of locally quartzitic sandstone; middle member of gray and green sandy shale; lower member of conglomeratic sandstone. Partly continental. Forms a ridge.
Upper Jurassic	Morrison formation	300	Shales of various colors containing thin sandstones and fresh-water limestones. Continental.
	Sundance formation	280	Upper part of green shale and shaly sandstone. Lower part of massive white sandstone grading downward into alternating white and red sandstones. Marine.
Triassic	Jelm formation	150	Coarse-grained thin-bedded maroon and white sandstones. Lenses of maroon conglomerate with fragments of limestone and red shale. Red shales and orange sandstones. Continental.
Permian	Chugwater formation	850	Brick-red shales and shaly sandstones. A limestone up to 10 feet thick, probably the Forelle, is locally present about 120 feet above the base. Massive gypsum beds in lower part. Mainly continental.
Pennsylvanian	Tensleep sandstone	300	Buff cross-laminated sandstones. Several thin sandy limestones near the base. Marine. Forms a ridge.
	Fountain formation	150	Red sandstones, shaly sandstones, white limestones, and pink arkosic grits. Partly marine.
Mississippian	Madison limestone	40	White massive limestone. Marine. Forms a ridge.
Cambrian	Unconformity	0–15	Local lenses of gray quartzite. Mapped as part of the Madison.
Pre-Cambrian			Pink potash granites, gray to pink injection gneisses, black and green schists, and metadiabases.

Amsden in the Rawlins uplift (Dobbin, *et al.*, 1928, p. 177). Pink arkosic grits in the upper part of the succession in the Elk Mountain district indicate that the beds here are, in part, equivalent to the Fountain formation of the Laramie Basin (Knight, 1929, p. 6-48).

The term Tensleep designates the same beds as in adjacent areas.

The redbeds between the Tensleep and Sundance in the northeastern part of the Hanna and Carbon basins were divided by Dobbin and associates (1929a, p. 11-12) into the Embar(?) and Chugwater formations. Their Embar(?) consists of 143 feet of red, green, and brown shales, gray limestone, sandstone, and gypsum overlain by a 3-foot lavender limestone, which they consider as probably corresponding to the Forelle limestone of the Laramie Basin. Their Chugwater, 1300 feet thick, consists predominantly of red shales, in part sandy, in which occur beds of sandstone, limestone, and gypsum; at the top is a zone 150 feet thick consisting almost wholly of sandstone.

Neely (1934, p. 5-6) applied the name Embar to the 300 feet of red shales, gypsum beds, and purple limestones immediately above the Tensleep in the southern part of the Elk Mountain district, the name Chugwater to the overlying 1000 to 1200 feet of red shales and shaly sandstones, and Jelm to the upper 75 feet of red sandstones and conglomerates. His total thickness is 1375 to 1575 feet, a figure considerably higher than that obtained by the writer.

Thomas (1934, p. 1655) states that marine limestone and sandstone tongues extending southeastward from the Phosphoria and Dinwoody formations of west-central Wyoming and interfingering with red shales in the base of the Chugwater in central and southeastern Wyoming constitute the "Embar" of central Wyoming. The Forelle limestone is one of the tongues of the Permian Phosphoria.

Knight (1917) recognized that in the Laramie Basin the upper 250 feet of redbeds, which here consist of sandstones and conglomerates containing limestone pebbles, clay balls, and fragmentary Triassic vertebrate remains, is separated from the underlying redbeds by a disconformity. For this upper part he proposed the name Jelm formation.

The writer found it necessary to subdivide the redbeds differently from other workers in immediately adjacent areas in order to obtain clearly recognizable lithologic units. In the southern and western parts of the district a limestone, probably the Forelle, is locally present about 120 feet above the Tensleep. It ranges in thickness from 10 feet to a foot, apparently pinches out in a number of places, and is not a satisfactory horizon marker for differentiation of the underlying red shales as Satanka. Other beds higher in the lower part of the succession are not sufficiently distinctive or persistent to make the Embar recognizable. The upper sandy and conglomeratic redbeds standing out as a ridge form a useful unit for mapping.

The term Chugwater, as used by the writer, includes the Embar(?), most of the Chugwater of Dobbin, and all of the Embar and Chugwater of Neely. The beds called Jelm are approximately equivalent to the Jelm of Neely and to the upper 150 feet of the Chugwater of Dobbin.

The formations from Sundance to Niobrara inclusive are the same as those of Dobbin and associates (1929a, p. 12-17).

The term Benton group, used in the same sense as by Darton and associates (1910, p. 9) in the Laramie Basin to include the Thermopolis, Mowry, Frontier, and Carlile, was adopted because in most places the shales are covered by debris

from the adjacent Cloverly ridge. Where adequately exposed, the formations were mapped separately.

Thomas (1936) has shown that the Carlile shale is of Niobrara age. The writer did not, however, include it in the Niobrara because, in most places, the Wall Creek sandstone immediately below is not well enough exposed for mapping.

In the Saddleback Hills quadrangle, which includes about 5 square miles of the northern part of the writer's map (Pl. 1), Dobbin and associates (1929a, Pl. 10) give as thicknesses of the Steele and Mesaverde 3600 feet and 2200 feet respectively. The writer obtained respective thicknesses of 2400 feet and 3500 feet. The columnar section by Dobbin and associates shows, beneath the base of their Mesaverde, sandstones and sandy shales 700 feet thick and beneath them a sandstone 500 feet thick. These beds were included by the writer in the Mesaverde in order to obtain units clearly recognizable by lithology and topographic expression. As used by the writer, the Steele consists of shales with a few thin sandstones and shaly sandstones, known as the Shannon member, 600 to 700 feet below the top; the Mesaverde consists of massive sandstone with thin shales and shaly sandstones.

The terms Lewis and Medicine Bow designate the same beds as in adjacent areas.

HANNA AND CARBON BASINS

Between the Medicine Bow and North Park formations in the Hanna and Carbon basins is a succession of sediments several miles thick. Lying upon the Medicine Bow in apparent angular accordance is the Ferris formation of which Bowen says (1918, p. 230-231):

"As defined the Ferris formation is approximately equivalent to the lower half of the 'Upper Laramie' of Veatch. The conglomerate at the base of the Ferris formation ranges through an interval of about 1,000 feet, in which massive sandstone, more or less conglomeratic, alternates with non-conglomeratic sandstone and shale. The conglomerate is sparingly developed at the base of the zone but increases in amount upward. It is succeeded by an alternating series of sandstone and shale containing numerous beds of coal. The maximum thickness of the formation is 6,500 feet.

"The Ferris formation contains fresh-water invertebrates, land plants, and bones of vertebrates. The shells and plants, which are found chiefly above the conglomerate zone, are regarded as of Fort Union age, whereas the vertebrate remains, thus far found only in the conglomeratic zone, consist of bones of turtles . . . and a few specimens that have been identified by C. W. Gilmore as *Triceratops*. The bones of *Triceratops* tend to show that the basal part of the formation should be correlated with the Lance of Wyoming and Montana and the Denver formation of the Denver Basin."

Of the overlying beds Bowen says (1918, p. 231):

"For the upper part of the 'Upper Laramie' as defined by Veatch the name Hanna formation is proposed. . . . The Hanna formation rests unconformably on the Ferris formation. A conglomeratic sandstone marks its base, above which conglomerate, sandstone, shale, and coal beds alternate to the top of the formation some 7,000 feet above the base.

"This formation differs from the Ferris in being highly feldspathic, in being conglomeratic throughout, and in the fact that the conglomerate contains an abundance of local material, notably granite, Mowry shale, and Cloverly conglomerate. . . .

"This formation contains an abundance of plant remains, all of which are referred to the Fort Union by Mr. Knowlton. The invertebrates are also regarded by Mr. Stanton as chiefly indicative of Fort Union age, though species near the top are said to resemble the Wasatch fauna."

The North Park formation rests unconformably on the Hanna and all older rocks (Bowen, 1918, p. 231).

TIME OF DEFORMATION

Stratigraphic relations in the Elk Mountain district show that deformation, other than possible slight warping, began after Medicine Bow time and before North Park time, an interval represented in the northern part of the Hanna Basin by some 13,000 feet of Ferris and Hanna beds. Little is known as to how much of this succession may have been deposited on, and subsequently eroded from, the area now occupied by the north end of the Medicine Bow Range. The problem hinges largely upon the question of the place from which the Ferris sediments were derived.

Bowen (1918, p. 234) presents evidence to show that the Ferris clastic material came from a distant source to the west. Knight (1937) concludes from conglomerate lenses in the lower Mesaverde of the northern Hanna Basin, and from conglomerates in the lower Medicine Bow in the embayment on the east side of the Medicine Bow Range (Fig. 1), that:

> "(1) Local anticlinal folds were raised above the Cretaceous sea prior to Mesaverde time, (2) the Cloverly formation was exposed on at least one of these folds at the beginning of Mesaverde time, (3) the pre-Cambrian basement complex was locally exposed at the beginning of Medicine Bow time, (4) the late Upper Cretaceous sediments having a total thickness of 15,000 feet in this region, were locally derived from the rising mountains, rather than from the remote Central Cordilleran geanticline, and were deposited in the adjacent basins."

It is generally agreed that the Hanna sediments were derived locally and that pre-Cambrian rocks were exposed at that time.

The occurrence of several thousand feet of nonconglomeratic Medicine Bow beds within 5 miles of the north end of Elk Mountain indicates little or no movement in the district at this time. One mile east of the northeast corner of the district the basal Hanna beds lie on Lewis shale (Dobbin, *et al.*, 1929a, Pl. 27), indicating that folding and erosion occurred in pre-Hanna time. Folding and erosion may have proceeded in the Elk Mountain district during at least part of the time when the Ferris conglomerate was being deposited. Dips up to 30° in the Hanna beds of the southern Hanna and Carbon basins indicate continuation of deformation after Hanna time.

COMPETENCE OF ROCKS

The rocks subjected to deformation consist of a complex of pre-Cambrian metamorphic and igneous rocks overlain by 17,000 feet of sediments including the Medicine Bow formation. Possibly there was

also several thousand feet of younger beds during the initial stages of movement. Stratigraphic relations of the Hanna formation indicate, however, a much smaller superincumbent load during the later stages.

The various types of pre-Cambrian rocks would necessarily deform differently under stress. Although the writer knows of no experimental work on mode of deformation of metamorphic rocks under subsequent stress applied in a direction different from that which produced the predominant metamorphic structure, it is probable that schists with excellent flow cleavage, under some conditions, deform by folding in much the same manner as well-stratified sedimentary rocks; gneisses show less tendency to slip along banding and are less subject to folding without faulting; and massive granites deform mainly by faulting or movement along closely spaced joints. In considering the influence of structural features of metamorphics during subsequent deformation, one must also take into account the probable orientation of later stresses with respect to cleavage, gneissic banding, faults, and contacts of massive igneous bodies in the metamorphics.

The 6600 feet of beds below the Mesaverde consists of shales and sandy shales with the exception of the massive limestones, grits, and cross-bedded sandstones in the lowest 500 feet of the succession, and the sandstones and conglomerates of the Jelm, Sundance, and Cloverly formations in the interval from 1350 to 2200 feet above the pre-Cambrian. These sandstone and conglomerate beds, all less than 100 feet thick, seem to have been of minor importance in transmitting stress and locally have acted somewhat differently from the shales mainly because of greater brittleness.

In the 275-foot interval above the Cloverly are numerous bentonite seams up to 2 feet thick. The writer believes that certain structural details described later were caused by ease of slipping along bedding in the bentonites, in contrast with difficulty of slipping in the coarsely cross-bedded Cloverly conglomerate.

The massive sandstones, up to several hundred feet thick, in the Mesaverde formation undoubtedly make this unit much more competent as a whole than the underlying beds. The shales and coals, however, permit bending without extensive fracture. The Mesaverde is the only unit in the area of tight folding with sufficient strength to lift at least part of its own weight from the crest of a growing anticline. This probably accounts for the common occurrence of "anomalous dips", apparently connected with drag folding, in the underlying Steele shale. It also explains, in part, the common difficulty of locating in depth the axial planes of unsymmetric sharp anticlines in the Steele shale and beds below.

DRAWING OF FOLDS

For parts of structure sections nearly normal to strike, surface dips were used as the primary control. Folds were drawn as concentric on the flanks, using as flank thicknesses the values given in Table 1. Crests of anticlines and troughs of synclines in the shale succession were rounded off; the few thin competent beds were drawn with constant thickness throughout. This procedure gives concentric folds in competent beds and a type of fold having the crest and trough thickening characteristic of similar folds in shales; it avoids the improbable relation of excessive flank thinning or even complete pinching out of beds required in the drawing of strictly similar folds with vertical or overturned flanks.

Because of changes of strike of beds in short distances, large parts of the sections are oblique to strike. For these parts the angles plotted were computed apparent dips. Vertical lines were drawn at points on the flanks near which surface dips are fairly constant for some distance normal to strike, indicating comparatively little variation of dip with depth. Along these lines were plotted depths computed from dip and flank thickness. Drawing across anticlines and synclines was carried out in the same manner as in cross sections normal to strike, except that flank apparent thicknesses were taken from a previously drawn part of the section controlled by depth lines.

The method outlined for drawing folds normal to strike of beds has been successfully used for many years by oil geologists in the Rocky Mountain region. The values they use for "thickness" in drawing beds that do not appear at the surface in an anticline are commonly obtained from exposures where the beds have a dip of 30° to 60° on the flank of a fold in the vicinity; the writer obtained the values for "thickness" in Table 1 by this procedure. Recent experimental work of Straley (1938) has shown that less competent beds in a succession of artificial sediments in a pressure box undergo thinning up to a 30-degree dip of limb, no thinning or even a slight thickening between 30° and 60°, and thinning from 60° to 90°. The value previously called "thickness" is therefore a close approximation to flank thickness for dips of 30° to 60°. As Straley's experiments were performed on materials for which strengths of competent and incompetent layers were known, his ratios of flank to crest thickness may be of doubtful applicability to a succession of rocks for which quantitative data on strength at the time of folding are not available. It is remarkable, however, how closely his ratios check with those derived from scalings of cross sections drawn by Bartram and Hupp (1929) across a number of sharp unsymmetric anticlines in the Rocky Mountains; in many cases correctness of drawing was later proved by drilling.

STRUCTURE

ELK MOUNTAIN

Elk Mountain (Pl. 1) is the eroded core of a strongly unsymmetric anticline. On the west side beds dip 15° to 30° W. The slope of the surface is only slightly less than the dip of beds, and consequently the resistant Madison, Fountain, and Tensleep extend far up the side of the mountain and form flatiron divides between deep gulches. On the east side the same formations dip at 65° or more and have reverse dips as low as 35°. The axial plane is closer to the east side, trends northeast at the south end, and a few degrees west of north at the north end.

East of the mountain base, pediment gravels and North Park sediments obscure structural detail. Between the pre-Cambrian and the few exposures of Cretaceous rocks, however, there is not enough space to accommodate the full succession of intervening beds; it is certain that there is a major fault here. The reverse dips of the Madison, Fountain, and Tensleep indicate that the fault is a thrust dipping west. On the basis of observed dips of the Bear Butte thrust farther west, in a similar structural environment, the dip of the Elk Mountain thrust is probably between 30° and 45°.

The thrust probably dies out southward in sec. 30, T. 19 N., R. 81 W., in the limb of an overturned anticline between the outcrop of Wall Creek sandstone in sec. 29 and the Mowry shale to the northwest. If the Wall Creek does not terminate against a fault beneath the alluvium or North Park, an improbably complex system of folds must be postulated. The fault between pre-Cambrian and Tensleep in sec. 9 of the same township is most likely a western branch of a multiple thrust fault. The horizontal distance of 4000 feet from the pre-Cambrian to the exposure of Steele in sec. 10 is sufficient to accommodate the full succession up to the base of the Steele with beds standing on edge. The 10-degree westward dip suggests, however, that the beds here belong to the east flank of a syncline and not to the east flank of Elk Mountain anticline. The multiple thrust consequently has a probable stratigraphic throw of several thousand feet. In the NW ¼ sec. 28, T. 20 N., R. 81 W., brown sandstones containing concretions up to 10 feet across undoubtedly belong to the lower part of the Mesaverde. The 60-degree dip away from the mountain front suggests upturning of beds in the block beneath the thrust. The stratigraphic interval of 6600 feet between the base of the Madison and the base of the Mesaverde, and the horizontal distance of 2500 feet from the Mesaverde exposure to the pre-Cambrian, indicate a stratigraphic throw of at least 4100 feet. Separation along the fault in section B′-B″ is 5500 feet.

At the north end of Elk Mountain the fault branches and changes strike from northwest to west. The high normal dips of beds near the branches in sec. 19 and the straight surface traces of faults across gulches indicate that the faults are nearly vertical and have the characteristics of tear faults rather than those of thrusts. The south branch clearly dies out in sec. 24. The north branch passes into nearly vertical shales of the Benton group in the NW ¼ sec. 19 and may continue westward to the Rattlesnake Pass fault. This would be difficult to demonstrate because the outcrop of the Benton group is in most places covered by debris from the Cloverly ridge, the width of the outcrop is sufficient to accommodate the full succession, and the fault, here nearly parallel to bedding, would produce little stratigraphic evidence of faulting even with large displacement.

The block between the two tear faults is cut by three transverse faults offsetting beds from Morrison to lower Frontier. The strike of the western fault is correctly represented, as exposures are fairly good and points within a few feet of the fault could be located for 1000 feet. Strikes of the other two faults may not be accurately represented. Both were mapped on the basis of a single observation of beds terminating along strike against younger or older beds; fault traces were drawn parallel to that of the western fault. Dips could not be obtained. Probably all three faults are genetically related to the apparent northeastward extension of the middle one.

North of the tear faults is a northward-pitching anticline, apparently the northward extension of the Elk Mountain anticline. Location of the axial plane could not be determined accurately enough to tell whether it is offset by faults or whether there are merely changes in trend. A fault striking northwest diagonally across the axial plane brings up older beds of the northwest flank. Dip and direction of dip of fault could not be obtained. If the fault dips southeast, it is normal and indicates elongation in this direction; all other features in the vicinity indicate shortening. If the fault is a vertical strike-slip fault, strike slip would have to be about 2500 feet to produce the offset shown by the Wall Creek and 1500 feet to produce that shown by the base of the Niobrara. Such large figures would require that the fault be propagated to the southwest and consequently offset the top of the Tensleep. The writer believes that the fault dips northwest as shown in section A-A′, and advances the following hypothesis to account for genesis.

The anticline developed under horizontal compression acting close to east-west; easiest relief was upward. Dips of beds increased until stratification in the shales and bentonites of the lower Benton group on the west flank was nearly parallel to the plane of maximum shear striking normal

to compression and dipping west. Stratification was also parallel to the shear couple set up by the bending of the shales over the crest of the anticline in the competent beds of the underlying succession from Jelm to Cloverly. Movement then began parallel to stratification. It produced a fault which was propagated downward parallel to stratification and upward across stratification east of the crest of the anticline from upper Benton through lower Steele. Such a fault would die out both upward and downward. Other cases of similar faults for which field data are much more complete are described later.

The writer's conceptions of movement of the Elk Mountain block are as follows:

The mountain is underlain by a scoop-shaped fault dipping at angles of 35° to 45° along the east side and close to vertical at the north end. The block above the fault moved relatively upward and eastward. The central or north-central part of the block moved farthest east, bending the front of the block into a curve convex to the east. The block as a whole was also rotated clockwise in plan, pivoting on a point somewhere near the south end of the mountain. Horizontal movement at the north end was, in part, by strike slip on both the steep northern part of the Elk Mountain fault and the steep Rattlesnake Pass fault.

NORTHEASTERN AREA

Simpson Ridge anticline.—The Hanna and Carbon basins (Fig. 1) are separated by Simpson Ridge anticline, also known as the Saddleback Hills anticline, for 12 miles north from the Elk Mountain district. General trend of the anticline is about N. 15° E. Formations as young as the Ferris cross the axial plane (Dobbin, *et al.*, 1929a, Pl. 27). Hanna beds dip away from the axis at angles up to 20° on both sides but do not cross the axial plane. For about 8 miles along the east flank the Medicine Bow and Ferris formations are overlapped by the Hanna, which lies on Lewis shale. Inasmuch as the Medicine Bow and Ferris are now exposed on both sides of the anticline at distances of 6 miles or less, the anticline has probably had the history, (1) folding after Ferris time, (2) erosion of the crest at least down to the Lewis, (3) deposition of Hanna, (4) folding involving Hanna beds, and (5) erosion to the present land surface. If this is true, the present trend is representative of that of pre-Hanna folding.

Simpson Ridge anticline extends south for 2 miles into the Elk Mountain district with the same trend of S. 15° W. In sec. 2., T. 20 N., R. 81 W., it pitches north at angles up to 50°. Southward trend changes to about S. 30° E. for nearly 2 miles. From here the fold trends southwest for 3 miles to the base of Elk Mountain. In secs. 13 and 24 the Shannon

sandstone of the east flank, judging from the poor exposures available, swings out to the east away from the outcrop on the west flank, indicating a dome, locally known as the Mill Creek structure, in secs. 13 and 14. The east flank is apparently the steeper. Various fold and fault complications are probably concealed beneath the alluvium. In order to work them out numerous pits, some many feet deep, would be required. The Midwest well in sec. 11 penetrated 1300 feet of Niobrara and yielded oil-saturated cores from the "Muddy" sandstone or Cloverly. The double thickness of Niobrara indicates drag folding, or faulting near the crest of the anticline.

From sec. 23 the anticline can be traced southwest into sec. 33. The few dips obtainable indicate that the southeast flank is the steeper. Shannon outcrops on the flanks projected beneath the alluvium are nearly parallel to the surface trace of the axial plane; as the land surface slopes from an elevation above 8000 feet in sec. 33 to about 7400 feet in sec. 23, the anticline pitches northeast. The thick cover of gravel and boulders, several feet in diameter in sec. 33, conceals relations of the anticline to the Elk Mountain block. The whole succession may change strike abruptly and be truncated by the thrust, or the anticline may continue beneath the fault and emerge as Pass Creek anticline. If this is true, the thrust has a displacement of several miles.

East from the south end of Simpson Ridge anticline is a gentle syncline trending southwest. The outcrops of Shannon sandstone dipping southeast in the NE ¼ sec. 26 and the SW ¼ sec. 23 indicate possible fault repetition on the northwest flank. The Shannon sandstone in sec. 24 outlines a syncline pitching southwest. The Shannon is repeated on the northwest flank by a high-angle fault striking east.

Halleck Creek syncline.—The Mesaverde forming the ridge in the northern part of the area outlines a sharp unsymmetric syncline with beds standing close to vertical on the west flank. Here and in the Lewis the fold pitches N. 20° E. at angles up to 20°. Within a mile beyond the north boundary of the map it dies out, as there is no appreciable swing in the outcrop of the Medicine Bow formation striking northeast from the west flank of Bloody Lake anticline to the west flank of Simpson Ridge anticline (Dobbin, *et al.*, 1929a, Pl. 27). South of the ridge the thick alluvial fill along Halleck Creek conceals the older rocks. It seems probable, however, that the syncline in secs. 15 and 16, shown by the exposure of a sandstone a few hundred feet above the base of the Mesaverde, is the same fold. Strikes of N. 60° E. strongly suggest that the syncline changes trend in the same manner as Simpson Ridge anticline. The Mesaverde exposed in the NW ¼ sec. 21 is the same bed, and

consequently the axis is close to the exposure. Previously described relations beyond the north edge of the map and the low flank dips near the mountain show that the central part of the syncline is the sharpest.

Bloody Lake anticline.—The map by Dobbin and associates (1929a, Pl. 27) covers a strip a mile wide at the north end of the Elk Mountain district. They gave the name Elk Mountain anticline to the fold in secs. 26 and 27. Structural relations to the south indicate considerable doubt that it is continuous with the anticline forming Elk Mountain. The writer consequently believes that the name Elk Mountain anticline should be reserved for the fold forming the mountain and has applied the name Bloody Lake anticline to the one in the northern part of the district.

The Mesaverde formation outlines a nearly upright anticline with flank dips of 60° to 80°. It pitches north at about 45° and dies out in this direction less than a mile from the map boundary. The Kasoming well in sec. 4, judging from its location with respect to exposures on three sides, started in nearly flat beds at least 1000 feet below the top of the Steele. On the basis of this assumption and normal thicknesses of formations, the "Muddy" sandstone should have been reached at 3400 feet. The driller's log shows that the well was abandoned at 4225 feet and did not reach the "Muddy" or Cloverly, both of which can be easily recognized from cuttings. Failure to reach beds below the Benton group can be explained by location of the well some distance off structure, or by deflection of the drill down one of the flanks, or by thickening of shales on the crest of the anticline.

A fault on the west flank produces a normal horizontal separation of 700 feet at the base of the Mesaverde. Northeast along strike sandstones in a zone up to 100 feet wide are offset by numerous steeply dipping fractures nearly parallel to stratification. Displacement decreases to the northeast, and the upper white massive sandstone member of the Mesaverde is not offset. Northwest of the fault the Lewis, Medicine Bow, and Ferris dip between vertical and 45° over a distance of 3 miles; the fault is therefore near the crest of the anticline. It is suggested that fracture and movement originated as follows:

The fold developed under nearly east-west horizontal compression and predominant upward relief. Sharp bending of the Mesaverde sandstones near the crest of the anticline produced a tension fracture when the flank dips of beds were lower than at present. The fracture was propagated downward toward beds lying nearly parallel to the westward-dipping plane of maximum shear. Stratification in the lower Mesaverde and upper Steele deflected the fracture to a position close to

the plane of maximum shear. During the later stages of deformation both the beds and fault were tilted more sharply to the northwest, producing the present high dips. The block northwest of the fault moved upward and was rotated on the curved fault counterclockwise as seen looking horizontally northeast, thus giving the prevailing higher dips of beds west of the fault. The northwest block might have moved horizontally northeast; this type of movement alone would account for the offset but would not adequately explain other observed relations.

Fort Halleck syncline and vicinity.—The syncline is named from the site of Fort Halleck near the northeast corner of sec. 20, T. 20 N., R. 81 W. A mile north from here the lower several hundred feet of the Mesaverde outlines a gentle upright synclinal basin. The surface trace of the axial plane, which can be fairly accurately located in several gulches, is convex westward. Dips of the lower Mesaverde in the SW ¼ sec. 16 show a change of trend and direction of pitch to southwest.

Between Fort Halleck and Halleck Creek synclines there must be an anticline, possibly the southward extension of Bloody Lake anticline. If this is true, Elk Mountain and Bloody Lake anticlines are different folds, as the axis of Elk Mountain anticline in sec. 20 is west of Fort Halleck syncline.

Eastward dips of the Steele in the SE ¼ sec. 7 indicate a sharp anticline between Fort Halleck syncline and the Mesaverde ridge to the west. If Bloody Lake anticline extends through this interval and is continuous with Elk Mountain anticline, it must undergo two improbably sharp changes of trend in secs. 17 and 18. Although the area in secs. 7 and 8 is undoubtedly generally anticlinal, it is most likely affected by a complex of faults, some of which are genetically connected with fracturing in the overlying Mesaverde, and others with deformation along Rattlesnake Pass fault.

Relation to Elk Mountain block.—In the area northeast of Elk Mountain only one fold, Simpson Ridge anticline, persists more than a few miles. It may terminate at the Elk Mountain thrust or may pass under it and emerge as Pass Creek anticline. In either case there is no direct structural relation between the anticline and the thrust block. The other folds farther west die out northward less than 7 miles from the mountain. They terminate southward at the thrust, or at the tear faults forming the western extension of the thrust, or in the belt of nearly vertical beds west of the tears. Shortening in the northeastern area was predominantly by folding; in the area farther south it was predominantly by movement along the Elk Mountain fault. Both blocks

were undoubtedly deformed, at least in part, during the same time interval, but the rocks yielded differently and more or less independently.

The structural pattern strongly suggests one produced experimentally in a pressure box by Link (1928). He applied uniform compression to a sheet of clay that had been strengthened by addition of cement in one corner of the box adjacent to the straight push block. A tear fault striking normal to compression developed between the areas of stronger and weaker materials. The weaker part was folded immediately in front of the push block; the stronger transmitted the stress.

Known lateral variations in lithologic character of the sediments of the Elk Mountain district indicate that the sediments had little lateral variation in strength. Observed relations north and east of Elk Mountain suggest that the rocks now constituting Elk Mountain were originally a more resistant mass in the pre-Cambrian basement complex analogous to the cement-strengthened mass in Link's pressure box.

Data on changes of trend of folds in the northeastern area are not entirely satisfactory because of poor exposures. The southern ends of several folds, nevertheless, seem to be offset east from their positions farther north. The following hypotheses are advanced to account for the offset.

1. Older folds trending north or northeast are affected by younger ones trending northwest. On the north side of the Hanna Basin (Pl. 2) major thrusts strike northwest. Folds northwest of the Elk Mountain district show the same trend; it appears, however, that some of them swing to south trend in the area west of Elk Mountain. The northwest-trending syncline in T. 22 N., R. 79 W., is definitely post-Hanna, as it affects beds of this age. Although these facts suggest a younger deformation producing northwest trends in the Elk Mountain district, they do not provide conclusive evidence of a younger cross folding intense enough to produce the indicated offsets of folds.

2. The S-shaped curves in surface traces of folds were produced by an eastward or northeastward movement of the sediments and basement rocks south of the curves after folds trending between north and northeast had been formed; movement was by bending of rocks or by strike slip along faults. The western part of the Elk Mountain fault shows this type of movement. Evidence is presented later to show that the block south of Rattlesnake Pass fault moved northeast.

NORTHWESTERN AREA

Rattlesnake Creek syncline.—The part of Rattlesnake Creek syncline west of Elk Mountain trends north. It is strongly unsymmetric; Triassic

and Jurassic beds in the east flank have dips as low as 30°, and the same beds in the west flank are overturned in many places. East of Bear Butte the syncline is cut longitudinally by a number of faults. Farther north it is cut diagonally by Rattlesnake Pass fault. Exposures here are not good enough to tell whether the axial plane is offset. In the lower Mesaverde north of Rattlesnake Creek the syncline pitches north at about 75°. Farther north pitch decreases, and the syncline passes beneath the unconformity at the base of the North Park. Trend here changes to a few degrees east of north, judging from strikes and dips of the Mesaverde some distance out on the flanks. The fold probably continues for 7 miles beneath the North Park and emerges as the gentle syncline shown by the Hanna beds in the southern part of the Hanna Basin (Pl. 2).

Sheephead Mountain anticline.—Bear Butte is part of the core of an anticline. The exposed rocks are principally pink massive granites, which are extensively fractured along the east border in the vicinity of Bear Butte fault. The pre-Cambrian rocks exposed at the south end of Sheephead Mountain are mainly schists, granite gneisses, and granites showing metamorphic structures that survived magmatic assimilation. Most of the mountain top is underlain by Madison limestone. From here north to Rattlesnake Creek sedimentary rocks form an anticline pitching north at angles up to 50°. Beds on the west flank have dips of 45° to 60°; the same beds on the east flank have been turned as much as 45° beyond vertical. Strikes and dips in the lower Mesaverde near the north boundary of the map show that the fold is here very sharp, nearly upright, and has a low northward pitch. The strike in the northern part of sec. 9 gives some indication that the axis swings to the northwest beneath the North Park. Sheephead Mountain anticline is probably the same fold as the northwest-trending Pass Creek Ridge anticline, which emerges from beneath the North Park 4 miles to the northwest (Pl. 2).

The Madison, Fountain, and Tensleep near the top of Sheephead Mountain are cut by four faults nearly parallel to the trend of the anticline. In all cases the west sides are upthrown. The eastern fault forms the boundary between Madison on the west, dipping 15° or less in this direction, and vertical or slightly overturned Fountain and Tensleep on the east. Along the line of section C-C′ horizontal Madison lies on the edge of vertical Tensleep. Farther south the Madison lies on various horizons in the Fountain. The fault is nearly horizontal and has a maximum net slip of several hundred feet. It was apparently formed by eastward movement of the plate of Madison limestone on

top of the anticline along the contact with the underlying pre-Cambrian and across the vertical beds of the east flank.

The two middle faults repeat the Madison and Fountain on top of the anticline and offset the top of the Fountain at the nose. The western fault, at its south end, repeats the Fountain and parts of the Madison and Tensleep. Displacement decreases northward, and only parts of the Madison and Fountain are repeated. The writer believes that the three western faults are high-angle thrusts or reverse faults dipping west at higher angles than the beds. Displacement is small, and consequently they may die out downward, as shown in the cross section, or may join the Sheephead Mountain thrust in depth.

Anticline east of Sheephead Mountain.—Formations from Jelm to Cloverly in secs. 33 and 34 outline the nose of a northward-pitching anticline. Southward the fold merges, in less than a mile, with the west flank of Rattlesnake Creek syncline. Three minor folds are superimposed on the anticlinal nose. These were probably caused by the competent beds in the succession between the Chugwater and Benton shales transmitting stress a short distance eastward from the Sheephead Mountain fault block. This conception was followed in drawing section C-C′; greater thickening of shales at crests of anticlines than in troughs of synclines produces minor folds that become less acute downward and merge into a simple anticline in the lowest beds.

Sheephead Mountain fault.—Sheephead Mountain fault, north of the junction with Rattlesnake Pass fault, brings Mowry shale into contact with Wall Creek sandstone. The beds here have a reverse dip of 45° W. Comparable reverse dips in other places, and the position of the fault between overturned anticline and syncline, indicate that it is a thrust dipping west at about 45°. North of the junction of the faults the Benton group outcrop widens, the Niobrara is not offset, and consequently the fault dies out in this direction. Southward older formations of the east flank of Sheephead Mountain anticline terminate against the fault. It then forms the boundary between this anticline and that to the east; the intervening syncline is faulted out. South of Hat Creek the Forelle limestone is offset only a few feet. The maximum stratigraphic throw of the fault is 650 feet, indicating maximum net slip of about 1000 feet.

Reservoir fault.—The town of Hanna obtains its water supply from a small reservoir located where Rattlesnake Creek crosses the outcrop of the Morrison in the SW ¼ sec. 26, T. 20 N., R. 82 W. A few hundred feet southeast the lower white sandstone of the Sundance is repeated by

a fault. At the line of section B-B′ Morrison on the west is upthrown against Mowry. The fault here dips 55° W. A few hundred feet north it branches and surrounds a slice of Thermopolis between Cloverly and Mowry. The fault then continues into the Mowry, and the beds in the east block are locally dragged up to a dip of 60° E. West of the center of sec. 23 the "Muddy" sandstone is offset only about 100 feet. The small offset and the lack of stratigraphic evidence of faulting in the northern part of the Mowry outcrop do not necessarily indicate small displacement, as here fault and beds closely approach the conditions of a bedding fault, a type that gives no stratigraphic evidence of faulting. Farther northwest, the Cloverly in the center of the nose of a southwest-pitching anticline is repeated. Reservoir fault terminates at Rattlesnake Pass fault.

In drawing the part of section B-B′ in the immediate vicinity of Reservoir fault, surface dips were adequate for the part to the east and for that 1500 feet west. Thicknesses and the 45-degree dip of the Niobrara near the center of sec. 27 give a control point at the base of the Madison well up the east flank of the syncline close to one located by depth from the Cloverly east of the fault. If the fault penetrates down to the Madison with the same dip as at the surface, the west block is downthrown in depth and upthrown at the surface. This indicates that the fault cuts across stratification in the competent beds in the succession from Jelm to Cloverly and flattens out parallel to stratification in the Chugwater shales. Branching of the fault around the slice of Thermopolis and relations in the Mowry at the surface indicate that the fault flattens toward stratification in this part of the succession also. The fault was apparently initiated by tension fracturing of sandstones at a place of sharp flexure and, under horizontal compression, was propagated downward and upward close to stratification in shales.

Rattlesnake Pass fault.—The Niobrara formation standing close to vertical on the west flank of Rattlesnake Creek syncline is offset 1000 feet by a fault whose strike is nearly normal to that of the beds. Southwest 1000 feet from the nearest outcrop of Niobrara the Cloverly is not offset; it is concluded that Rattlesnake Pass fault terminates against Sheephead Mountain thrust. The lower Niobrara beds on the east flank of the syncline are exposed 1200 feet southeast of the base of the Mesaverde, indicating a stratigraphic throw of 1900 feet. At the crest of Rattlesnake Pass in the NW ¼ sec. 23, where beds and fault have the same strike, exposures of lower Benton within 500 feet of the base of the Mesaverde show that the fault has a stratigraphic throw of 4000 feet. In the SE ¼ sec. 14 at least 1600 feet of Steele shale is faulted out.

In sec. 13 the outcrops of nearly vertical Shannon and Niobrara diverge eastward providing, in the eastern part of this section, sufficient space to accommodate the full thickness of intervening beds. Dip of the fault could not be obtained. The fairly straight surface trace indicates that it is nearly vertical; dips of beds in the vicinity give no contrary indication.

Rattlesnake Creek syncline is clearly present on both sides of the fault, although exposures are not good enough to determine whether its axial plane is offset. Absence of offset of Mesaverde beds in the NW ¼ sec. 23 shows that Reservoir fault terminates at Rattlesnake Pass fault. The sharp northwest-trending anticline in sec. 24 swings, farther west, parallel to Rattlesnake Pass fault; no anticline is found on the other side. The fact that some structural features terminate at the fault and cannot be matched on the other side indicates that Rattlesnake Pass fault is of rotatory type (Reid, *et al.*, 1913, p. 167). If small parts are considered as translatory, a fairly accurate conception of the rotatory movement as a whole can be obtained.

On the west flank of Rattlesnake Creek syncline, the Niobrara beds and Rattlesnake Pass fault approach conditions of vertical beds cut by a vertical fault. A large dip slip, in this case, produces no separation. Horizontal separation parallel to strike of fault is equal to strike slip. It is concluded that here the Niobrara in the southeast block moved horizontally northeast about 1000 feet, the observed horizontal separation parallel to strike of fault. In the NW ¼ sec. 23 strike of fault and beds are the same. Strike slip on a strike fault produces no omission or repetition of beds; these are produced by dip slip only. On the basis of a vertical fault, observed omission of 4000 feet of beds, and 50° as the mean dip of the beds, the dip slip here is around 6000 feet. Relations in the SE ¼ sec. 14 do not provide adequate data for obtaining approximate values for net slip or any of its components. Decrease in stratigraphic throw northeastward from sec. 23 indicates, however, that net slip probably decreases in this direction.

Underground relations along Rattlesnake Pass fault are believed to be approximately as shown in sections A-A′ and B-B′-B″. The dotted lines below ground east of the vertical fault in section B-B′-B″ represent the trace on the vertical fault of stratification in the north block; the trace is projected horizontally south onto the plane of section. The dotted lines to the west represent the trace on the vertical fault of the base and top of the Madison in the south block; the trace is projected horizontally north onto the plane of section. Rattlesnake Pass syncline has a low pitch, and consequently relations in depth as shown differ only slightly from those that would appear on a cross section in the plane of the fault. The section as drawn has the advantage of showing

the vertical fault terminating upward and dying out westward against Sheephead Mountain thrust. Quantitative relations along the vertical fault in section A-A′ are probably inaccurately represented because the line of section for more than a mile to the east is nearly parallel to strike of beds with high normal and reverse dips. The writer believes that relations were produced as follows:

After Rattlesnake Creek syncline had been slightly folded a vertical fracture was formed. As the syncline became sharper, the anticline in secs. 23 and 24, and Reservoir fault were initiated. The block south of Rattlesnake Pass fault moved up to the east along Elk Mountain fault and along the vertical faults north of the block. While the south block moved as a whole, the anticline in sec. 23 was sharpened, dragged around to its present position, and cut by Reservoir fault. Later stages of deformation in the north block merely produced sharpening of the Rattlesnake Creek syncline. At some time, probably in a late stage, Sheephead Mountain thrust block moved east over the west end of Rattlesnake Pass fault.

The mode of origin of Rattlesnake Pass fault, a vertical fracture striking at about 45° to trend of major folds, is considered in the following hypotheses:

(1) The fracture coincides with a plane of weakness in the pre-Cambrian rocks such as a fault, stratification, schistosity, or an intrusive igneous contact. The plane of weakness forms the boundary between more competent crystalline rocks to the south and less competent rocks to the north. Under east-west compression, a fracture developed in the basement rocks and was propagated upward into the sediments. In the core of the Medicine Bow Range 30 miles southeast is a large area of metamorphics described by Blackwelder (1926). A summary of his description has been given in the preceding section on pre-Cambrian rocks. Bedding, schistosity, and a major fault strike northeast and dip steeply to the southeast. In the northern part of the Laramie Range foliation in schists strikes northeast, and contacts between schists and intrusive rocks are as a rule steeply inclined or nearly vertical (Spencer, 1916, p. 52). Therefore, a plane of structural weakness in the pre-Cambrian at Rattlesnake Pass fault is not improbable.

(2) The vertical fracture was formed under east-west compression with easiest relief horizontally to the north. Under these conditions the planes of maximum shear are vertical and strike respectively northeast and northwest. As previously pointed out in the section on time of deformation, it is generally accepted that the pre-Cambrian cores of rising mountains adjacent to the Hanna Basin were exposed during

Hanna time. Bowen (1918, p. 234) and Knight (1937) disagree as to whether the mountains began to rise at an earlier date. If, however, at some time during deformation under east-west compression, Elk Mountain stood out in relief above an adjacent basin to the north which was sinking under a load of new, poorly consolidated sediments, lateral relief to the north would be afforded. This hypothesis does not conflict with the first one; a vertical plane of weakness striking northeast in the pre-Cambrian basement would favor development of a fracture under east-west compression with easiest relief horizontally to the north.

Bear Butte fault.—In secs. 4, 10, and 23, T. 19 N., R. 82 W., Bear Butte fault crosses topography rugged enough to make it possible to obtain dips, which are from 35° to 45° W. The fault is a westward-dipping thrust. Near the center of sec. 4 it passes beneath Tertiary sediments. To the south in secs. 10 and 15 the branches numbered 1 and 2 surround a crescentic slice of rocks from pre-Cambrian to Chugwater in age, which belong to the west flank of Rattlesnake Creek syncline. Beds here are vertical or overturned with dips as low as 35° W. West of the center of the slice is a block of Madison, 1000 feet long and up to 150 feet wide, surrounded in plan by pre-Cambrian. The block could have reached its present position in a number of ways. The relation that seems simplest is shown in section D-D′. Fault 1 cut off a slice of Madison from the crest of a flexure below the present land surface and carried it upward onto the pre-Cambrian.

The high ridge formed by the Cloverly of the west flank of Rattlesnake Creek syncline in sec. 10 terminates abruptly in the south-central part of the section against beds of the Benton group. East of the south end of the ridge Thermopolis shale of the west flank of the syncline is exposed west of Frontier shale of the east flank. There are no intervening exposures of Mowry. The Mowry is probably absent, as it usually stands out in relief or occurs as silver-gray angular fragments in deeply weathered material covering the outcrop. Indications are that a thrust here cuts the Benton shales near the axis of the syncline. West of the south end of the ridge blocks of Jelm and Cloverly conglomerates up to 10 feet across are found along the outcrop of thrust 2 as far south as the outcrop of Cloverly in the north-central part of sec. 15. The blocks indicate that the Cloverly extends from the outcrop in sec. 15 beneath thrust 2 and is brought out to its position in sec. 10 by tear fault 5. This fault does not offset the Tensleep of the west flank of Rattlesnake Creek syncline or the Cloverly of the east flank. It is therefore concluded that fault 5 terminates upward at thrust 2 and downward at thrust 3.

Cross section D-D′ shows relations south of fault 5. A section a short distance north would be similar but would probably show, near the axial plane of the syncline, thrusting of younger beds onto older. This relation is found about 6 miles south (section L-L′) and also at several places in the southern part of the Medicine Bow Range (Beckwith, 1938).

Northeast of fault 4 the beds from Jelm to Cloverly belong to the vertical or slightly overturned west flank of Rattlesnake Pass syncline. In the block to the southwest, bounded by faults 4, 3′, 2, and the southward extension of 2 and 1, beds from Chugwater to Morrison dip northwest about 35° and are progressively younger to the northwest; the block obviously belongs to the east flank of the syncline. Fault 4 makes a straight trace across a steep gulch 150 feet deep and is therefore nearly vertical. It does not offset the Fountain to the northwest or the Cloverly to the southeast, indicating that it is a tear fault terminating upward at thrust 2 and downward at thrust 3′. The surface trace of fault 3′ trends northeast in the area of low relief near the bottom of the gulch and then swings sharply to the southeast up the side of a spur whose crest descends sharply to the northwest along line F-F′. Evidence indicates that fault 3′ near the surface is a thrust dipping northwest at an angle greater than the dip of the beds.

General relations shown in cross section F-F′ are similar to those in D-D′. The main difference is that in F-F′ the beds are overturned only in the extreme western part of the syncline. In cross section E-E′, which crosses tear fault 4, the eastern dotted line represents the trace of thrust 3′ on vertical tear 4; the trace is projected horizontally north onto the plane of section. The western dotted line represents the trace of thrust 3 on tear 4 projected horizontally south onto the plane of section. Relations shown suggest that thrusts 3 and 3′ are probably a single fault in depth and that folding north of the tear was independent of movement of rocks immediately to the south. Fault 4 is of marked rotatory type.

In the SW ¼ sec. 15 and the NW ¼ sec. 22 Bear Butte fault cuts southeastward across the axis of Rattlesnake Pass syncline and brings up older beds on the southwest. At the boundary between secs. 22 and 23 the Madison, Fountain, and Tensleep of the east flank of the syncline change strike abruptly and pass under an east branch of the thrust. They are brought up from beneath the pre-Cambrian east of the center of sec. 22 as slice blocks along a multiple thrust. Relations shown in cross section G-G′ are explained as follows: The two northeast faults near the surface dip at angles slightly greater than the beds. In depth below the junction of these two, dip of fault is less than that of beds,

and the pre-Cambrian moved northeast across a small syncline whose core is formed by Tensleep. The Fountain and Tensleep in normal succession between the two southwest faults were carried up from the syncline along faults dipping more steeply than the beds. The overturned Madison and Fountain farther southwest form part of the overturned southwest flank of the syncline.

In the SW ¼ sec. 23 a block of Tensleep 1200 feet long and 300 feet wide is completely surrounded, in plan, by pre-Cambrian. The sandstone is extensively brecciated and folded, but the predominant direction of dip is southwest. Probable relations shown in section H-H′ are, in general, closely similar to those in section G-G′. The normal fault shown in sections H-H′ and F-F′ has no relation to thrust faulting; it is a tension fracture formed by bending near the crest of Elk Mountain anticline. In the northern part of the slice block of Tensleep 50 feet east of the west fault there is an abandoned mine shaft 50 to 75 feet deep, judging from the size of the dump. All of the material is extensively brecciated Tensleep sandstone; a few fractures contain small amounts of malachite and chalcocite. Mineralization is definitely not genetically connected with igneous activity, as the nearest igneous rocks, other than pre-Cambrian, are on the west side of the Sierra Madre Range a few miles from the State line. Probably hot meteoric waters circulating along the fault zone at considerable depth dissolved small amounts of copper from pre-Cambrian metadiabases and hornblende schists and deposited it in the Tensleep.

Bear Butte thrust shows unusual structural relations. On the east side of the butte it occurs in the limb between overturned anticline and syncline in a common structural association. Southeast of Bear Butte it cuts diagonally across Rattlesnake Creek syncline and Elk Mountain anticline. Relations described later show that it dies out to the southeast in the limb between overturned anticline and syncline on the east side of Coad Mountain. The pre-Cambrian rocks exposed in Bear Butte are massive granites; those exposed near the thrust branches in secs. 22 and 23 are metadiabases and hornblende schists. The rocks of the southern part of Elk Mountain and the northern part of Coad Mountain are massive granites containing a few tabular remnants of metadiabases. Transgression of the central part of the thrust across axes of major folds might be explained by fault movement under northeast-southwest compression after the formation of folds under east-west compression. This hypothesis fails to explain why the northern and southern parts of the thrust strike parallel to trend of folds. The writer believes that deflection across trend was caused by differences in strength and structure of the rocks in the pre-Cambrian basement complex.

COAD MOUNTAIN AND PENNOCK MOUNTAIN AREA

Bear Butte fault and vicinity.—In the northern part of sec. 26, T. 19 N., R. 82 W., Bear Butte fault passes under the North Park formation. A steep gulch entering Pass Creek from the south near the center of the section is apparently cut along the trace of the fault. At the head of the gulch there is a saddle in the granite ridge that descends toward the creek. To the southeast steeply dipping beds from Madison to Chugwater on the east flank of Elk Mountain anticline terminate against the fault. In secs. 25 and 36 beds from Chugwater to Frontier outline a southward-plunging overturned anticline whose axial plane dips west. West of the anticline a fault dies out northward in the east flank of Elk Mountain anticline and joins Bear Butte fault to the south. At the junction of the faults pre-Cambrian of the east flank of Coad Mountain anticline lies on Thermopolis of the west flank of the small overturned anticline. This relation was apparently produced by movement of Coad Mountain anticline eastward across an intervening syncline as shown in section J-J'. A minimum value for the combined net slips of the two faults at this place is 2500 feet. The block of sediments from Madison to Chugwater between the two faults is part of the west flank of the syncline.

In the SW ¼ sec. 36 a branch of the thrust brings Cloverly of the east flank of Coad Mountain anticline onto Frontier of the south nose of the small overturned anticline. From here south fault displacement decreases rapidly. North of the center of sec. 1 the full thickness of the succession from Morrison to Niobrara is present with the exception of about 400 feet of Frontier. East of the center the short interval between exposures of overturned Mowry and Niobrara indicates faulting out of a few hundred feet of beds. The thrust probably dies out in overturned Steele shale in sec. 12. The Cloverly southeast of Coad Mountain is exposed in enough places so that one can trace the outcrop beneath cover. It is affected by several small folds pitching south or southeast. In the S ½ sec. 1 there is a sharp overturned anticline and a gentler syncline to the west. The short interval along the north boundary of this section from the outcrop of the Madison, projected beneath the North Park, to the Cloverly shows the presence of a branch of Bear Butte thrust, which probably dies out in the east flank of the overturned anticline. Changes of strike and dip of the Cloverly in the NW ¼ sec. 12 suggest another minor anticline.

Southwestern part of Coad Mountain.—Beds from Madison to Chugwater show, in secs. 3 and 10, an anticline pitching at a low angle toward Pennock Mountain. Trend of the northern part is a few degrees west of south, and that of the southern part a few degrees east of south. In secs. 2

and 3 the crest of the anticline is cut by six faults striking roughly parallel to trend of fold. Their straight surface traces indicate that the faults are nearly vertical. Offset and elevations of the base of the Madison in contact with the west fault indicate a throw of about 400 feet; throw of the others is less. In all cases the west block is downthrown. Occurrence of the faults in brittle beds and granites on the crest of an anticline, and the fact that the faults strike nearly parallel to its trend, indicate that they were formed as tension fractures during folding. It is shown later that a thrust sheet advanced from the west over the southern part of the anticline (sec. L-L′). Although the North Park obscures the structure of the older rocks west of the tension-faulted anticline, it was probably buried beneath the sheet. Such a sheet advancing eastward would successively load and depress the block west of each of the tension faults.

Coad Mountain fault.—Beds in the E ½ sec. 11 outline a gentle anticlinal nose pitching about 40° SE. Coad Mountain fault here cuts the beds nearly normal to strike. Farther north it strikes N. 25° E. parallel to strike of beds in the east block and at 45° to strike of beds in the west block. The fault trace, in crossing the gulch at the road, is convex westward, indicating that the fault possibly dips west at a low angle. If this were true, the trace south of the ridge formed by the Jelm in the east block should swing west in approaching the valley to the south. Curvature of the trace is therefore the result of a change of strike, not of erosion. The fault plane is nowhere well exposed, but, looking north along strike across the deep gulch at the road, it appears to have a dip between vertical and 60° W.

Maximum stratigraphic throw, about 800 feet, is shown where Madison in the east block is in contact with middle Chugwater. From here southeast, throw decreases, and the fault probably dies out in sec. 12. In sec. 2 the high normal and reverse dips of the Madison east of the fault suggest that it is a westward-dipping thrust beneath which there is a syncline containing beds as young as Chugwater. This relation is shown by Neely (1934, Pl. III, sec. G-H). The writer does not believe that this is true because the beds are not overturned west of the fault, and there is no independent evidence of a syncline along the strike of the fault. Evidence indicates that Coad Mountain fault is normal, as shown in section K-K′. The following hypothesis is given to account for relations shown.

During an early stage of deformation, Coad Mountain fault originated as a nearly vertical tension fracture near the crest of an anticline. Bear Butte thrust was later formed and cut Coad Mountain fault. The

sedimentary rocks between the Tensleep and Mesaverde are almost entirely of types that bend easily; it is fairly certain, therefore, that the thrust was initiated in the pre-Cambrian and was propagated upward to the east into the sediments. Beds were warped into a steep monocline in front of the edge of the growing fracture. As the fracture penetrated eastward into the monocline, stratification in vertical to overturned beds deflected the fracture upward parallel to bedding, producing a thrust that steepens toward the land surface in the sediments and has a fairly constant dip in the pre-Cambrian. As shown in section K-K′, the block between Coad Mountain fault and Bear Butte fault is a wedge bounded below by a surface concave upward. Eastward movement of both blocks above the thrust would cause the wedge to move up with respect to the blocks on both sides. Shortening in the upper part of the wedge consisting mainly of sediments would give the beds a higher dip to the east, give Coad Mountain fault a lower dip to the west, and cause its footwall to move up. Southward change in strike of Coad Mountain fault is believed to be the result of movement on Pennock Mountain fault.

Pennock Mountain fault system.—In the SW ¼ sec. 10 the alluvial deposits have been stripped from the crest of a knob. From east to west there are exposed (1) Chugwater, (2) a fault omitting 1000 feet of beds, (3) Mowry shale overturned and dipping 50° SW., (4) Thermopolis, and (5) Cloverly. The Chugwater is in normal position on the west flank of the anticline extending south from Coad Mountain. The other beds become progressively older westward and consequently do not belong to the west flank of this anticline; they are part of the overturned east flank of the anticline whose pre-Cambrian core forms the ridge at the northwest end of Pennock Mountain. West of the center of sec. 10 Cloverly and Thermopolis of the east flank of the west anticline are exposed within 600 feet of Tensleep of the west flank of the east anticline. At least 1700 feet of beds is cut out by a fault beneath the valley fill. Evidence indicates that the west anticline moved eastward across a syncline and onto the east anticline as shown in section L-L′.

Normal to strike the distance from the westernmost Cloverly exposure to the pre-Cambrian is 1200 feet. The stratigraphic interval between them is 2080 feet. There must be, therefore, a fault between pre-Cambrian and Cloverly outcrops. It probably joins the other thrust in depth as shown in section L-L′.

The value for net slip obtained by scaling from a cross section is dependent upon accuracy of the section. Even with complete surface data the syncline beneath the thrust could not be drawn accurately. Minimum net slip is obtained if it is assumed that the syncline is closed

and cut by a dip-slip strike thrust normal to the axial plane of the syncline. An open fold or a smaller angle between thrust and axial plane of syncline gives a greater value for net slip. In drawing section L-L′ the assumptions were made that the fold is closed, the thrust has about the same dip as other observed thrusts cutting the pre-Cambrian in the vicinity, and that the beds on the east flank of the syncline have a maximum dip of 65°. These assumptions give a conservative result. Separation along the lower thrust shown by the base of the Mowry is 4700 feet. Combined net slip for the two thrusts is considerably greater than a mile.

A possible indication of much greater displacement on the thrust is found near the center of sec. 11. Here a mass of pre-Cambrian rocks 200 feet across forms the crest of a knob. The contact with the underlying Mowry shale is nearly flat. At first glance the pre-Cambrian appears to consist of boulders similar to those in the lower part of the North Park formation. Many of the blocks, however, show no rounding, and tabular masses of metadiabase and pegmatite dikes in the gneisses can be traced laterally for 50 feet. It is concluded that the pre-Cambrian is a brecciated mass that reached its present position by movement as a whole. The mass is likely a klippe lying on part of the Pennock Mountain thrust, in which case net slip of the fault is several miles. An alternate explanation is that the pre-Cambrian block came from the south by creep and landslide at a time when the north face of Pennock Mountain stood out in higher relief than at present. Evidence against this hypothesis is the horizontal distance of 3000 feet from the block to the scarp, probable high dip of the fault at the base of the scarp, and the fact that the beds dip toward the main pre-Cambrian mass, thus presenting the most difficult possible conditions for the development of northward-dipping landslide faults.

The lower thrust has the characteristics of an inclined subsequent shear thrust (Billings, 1933, p. 147) in that it cuts across a fold at an angle less than the dip of axial plane and produces both thrusting of older rocks onto younger and thrusting of younger onto older. The writer believes, however, that the genesis implied by the term may not fit the case of Pennock Mountain thrust. Observed relations could have been produced as follows: (1) A break thrust cut the west limb of the syncline; (2) erosion reduced the surface of the east block nearly to its present level, or even lower near the thrust; and (3) the west block advanced over the erosion surface. Stratigraphic relations in the Hanna Basin previously described in the section on time of folding show that in adjacent mountain areas there were alternate intervals of deformation and

erosion and that by Hanna time pre-Cambrian rocks were exposed. This history fits the conception that Pennock Mountain thrust is genetically a combined break and erosion thrust.

That the north end of Pennock Mountain is bordered by a fault is shown by exposures of Paleozoic and Mesozoic rocks closer to the pre-Cambrian than their respective stratigraphic distances above the pre-Cambrian. Neely (1934, Pl. III) shows a thrust continuing south for 8 miles along the east side of the mountain. The writer's conceptions of structure, movement, and genetic relations of the Pennock Mountain block are closely similar to those previously expressed for the Elk Mountain block and are summarized as follows: Pennock Mountain is underlain by a scoop-shaped fault dipping at angles of 35° to 45° along the east side and close to vertical at the north end. The block above the thrust moved relatively upward and eastward, and deformation within its northern part was more or less independent of deformation within the southern part of the Coad Mountain block. The rocks constituting Pennock Mountain block were originally a more resistant mass in a heterogeneous basement complex.

Evidence of high dip of fault at the north end of the mountain is shown in the southern part of sec. 11. Exposures of Wall Creek and Sundance are closer together than the thickness of intervening beds and dip south at an angle greater than 40°. A southward-dipping fault producing repetition of the Sundance and Morrison must have a dip greater than that of the beds, as shown in section M-M′. Eastward movement of the Pennock Mountain block is indicated by the eastward swing of Coad Mountain fault and the small anticline at the southwest end of Coad Mountain as they approach the tear fault.

Relations of the tear to the two thrusts west of the small anticline are obscure because of extensive alluvial cover. South of the eastern thrust a ridge rises in this direction toward the top of Pennock Mountain. Therefore, the thrust probably terminates laterally against the tear. South of the inferred position of the western thrust there is a deep valley in the pre-Cambrian; it is therefore likely that the thrust continues southward in the pre-Cambrian. On the crest of the pre-Cambrian ridge west of the deep valley a saddle on the strike of the tear indicates that the fault crosses the ridge. In the SW ¼ sec. 10 it is possible that, beneath the alluvium, the tear and western thrust intersect without one offsetting the other or with offset much less than would be expected from inferred net slips of faults. Absence of offset is, however, the relation that should occur, inasmuch as inferred directions of movement on both faults are nearly parallel to the line of intersection of the fault surfaces. A similar

case of intersection of thrust and tear farther south in the Medicine Bow Range has been mapped (Beckwith, 1938, Pl. 1). An area of excellent exposures in sec. 26, T. 14 N., R. 77 W., shows a major thrust offset only several hundred feet by a tear having a net slip of at least 2000 feet. Beds here, however, show large offsets and omissions. Beds would show the usual stratigraphic criteria of faulting in the vicinity of thrust and tear intersection only when stratification parallels one of the faults and makes a trace on the second parallel to direction of movement on the second.

PASS CREEK BASIN AREA

In secs. 29 and 32, T. 19 N., R. 81 W., the Wall Creek sandstone outlines a north-trending open syncline pitching both north and south away from a point near the boundary between the sections.

Pass Creek anticline is a sharp fold trending north for at least 4 miles. It may continue south beneath the terrace deposits in sec. 8 and north beneath Elk Mountain thrust. The anticline is markedly overturned; the Wall Creek dips 30° to 55° on the west flank and is overturned as much as 40° beyond vertical on the east flank. East of the center of sec. 5, where the anticline is not faulted, it is a closed fold. A lenticular sandstone in the Frontier of the west flank dips 50°; on the east flank only 200 feet away the sandstone is overturned, as shown by cross lamination, and has the same dip. The pitch is low. Frontier is exposed in the core in the southern 3 miles except where the Mowry is brought to the surface by faults. The exposures in sec. 20 show that Carlile here crosses the axial plane. Although these exposures are 300 to 400 feet higher topographically than those of Mowry and Thermopolis to the south, the anticline definitely pitches north at a low angle.

In sec. 32 on line J-J′ full thickness of Frontier is present on the west flank, and approximately half is faulted out on the east flank. The Mowry in the core dips uniformly west at 40°. Its east contact is well exposed in transverse gulches up to 15 feet deep. The lower few inches is extensively brecciated and recemented along a fault dipping 40° W. parallel to bedding in the Mowry and cutting across the edges of Frontier beds, which are vertical to slightly overturned. The fault is a bedding thrust dipping toward the gradual flank of the fold, as shown in section J-J′. It probably dies out in the Mowry to the west and undoubtedly offsets the axial plane of the anticline. The restoration above ground shows the relation of decrease in slip upward to the east, where the fault plane cuts at progressively increasing angles across stratification and dies out in a sharp flexure.

The fault is genetically connected with differences in ease of slipping along stratification. The succession below it from Jelm to Cloverly con-

tains massive sandstones and cross-bedded conglomerates; the Mowry is thinly laminated and contains numerous bentonite seams. Stratification in the west flank of the anticline coincides with one of the planes of shear of regional horizontal compression and also with the local shear couple set up by bending of beds over the anticline. The lower shearing strength of the Mowry parallel to stratification caused initiation of movement at this horizon on the west flank. This was followed by propagation of a fault downward to the west toward the syncline and upward across the crest of the anticline.

The exposures of Mowry and Frontier in the NE ¼ sec. 29 indicate that there is a westward-dipping thrust cutting across the crest of Pass Creek anticline. The Mowry shows only westward dips; a few feet to the east the Frontier is nearly vertical.

In secs. 8, 9, and 10 the Carlile and Niobrara outline three small anticlines and two synclines. They pitch northwest, and their axial planes dip at high angles to the northeast. In the N ½ sec. 8 the western exposures along the creek show beds from Niobrara to Mowry that belong to the west flank of Pass Creek anticline. From 100 to 200 feet of the lower Frontier is faulted out. In further discussion and in the cross section this feature is neglected, as it fits in genetically with major features and is too small to draw on the scale used. The slice of Mowry is in fault contact to the east with Niobrara of the southwest flank of a small anticline. Stratigraphic throw of the fault is at least 1000 feet. Its dip could not be obtained. Probable relations as shown in section K-K′ are similar to those in section J-J′ except that displacement is greater. The minor folds east of the thrust indicate that some of the beds in the sedimentary succession were capable of transmitting stress several miles from the vicinity of the thrust. The competent beds from Jelm to Cloverly probably glided over the Chugwater shales and gypsums and underwent shortening by folding east of the thrust. Greater eastward or less westward movement of the competent beds as a whole in relation to the overlying incompetent beds would give folds whose axial planes dip east down to the Chugwater.

The fault in sec. 33 was formed by a landslide. The edge of the North Park forms a steep scarp above Steele shale. The block west of the fault slipped down and was rotated clockwise as seen looking horizontally north. The dips shown were caused by rotation and drag along the fault.

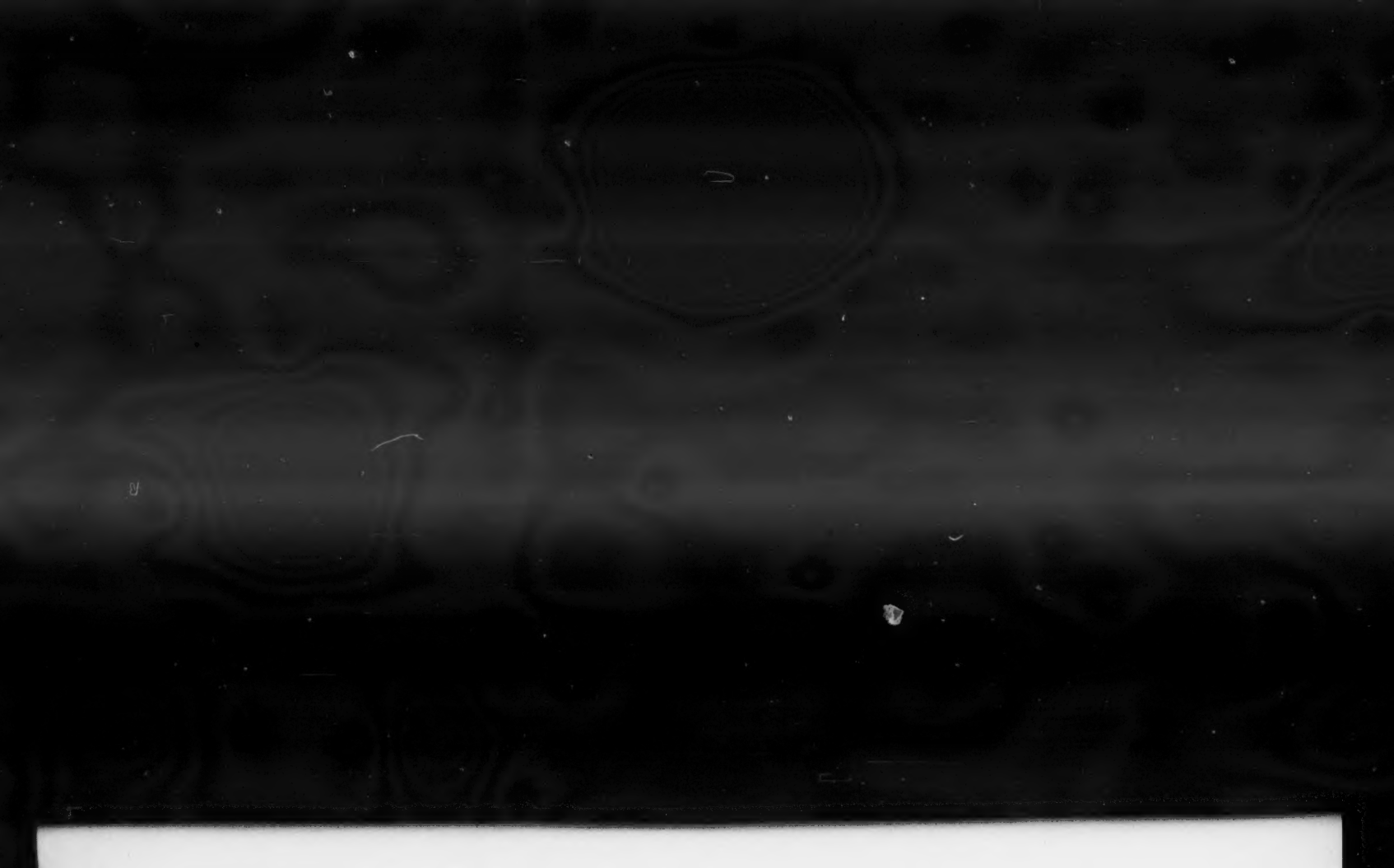
OIL AND GAS POSSIBILITIES

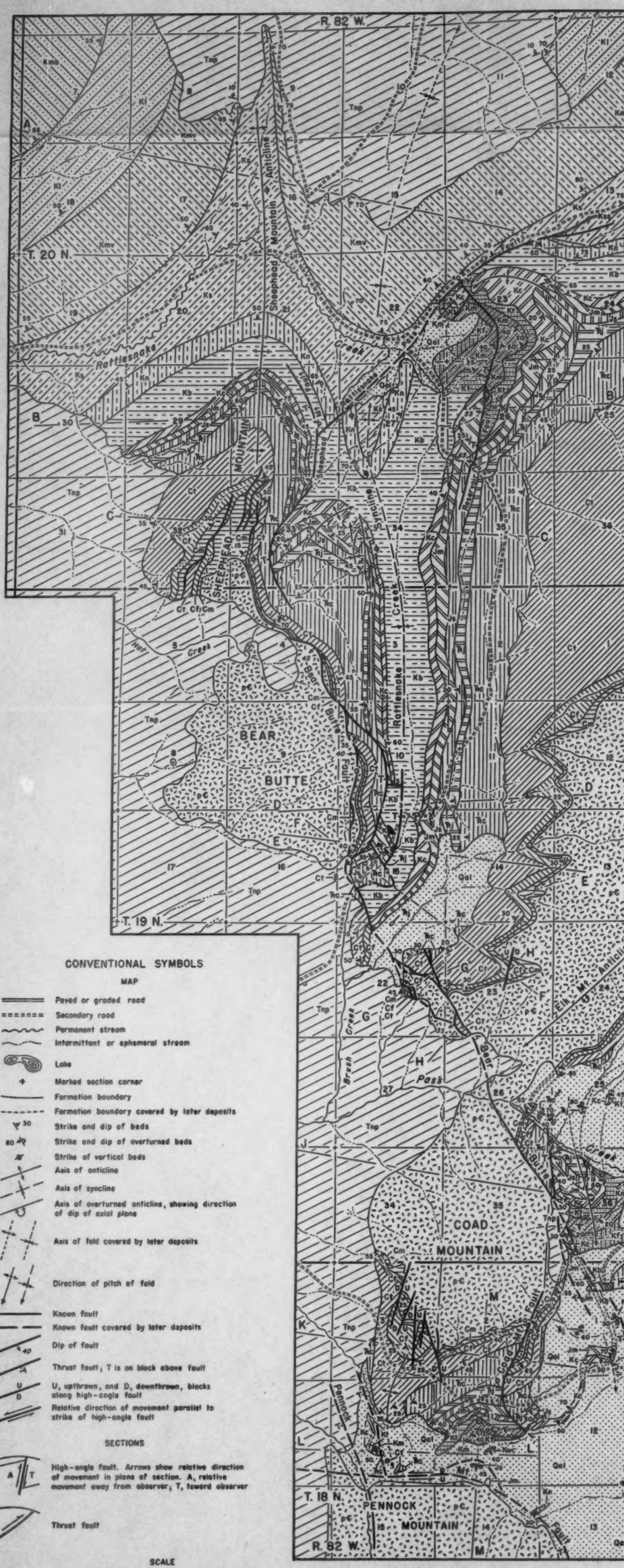
General statement.—Possible occurrence of oil and gas in commercial quantities in the Elk Mountain district is largely a matter of structural

traps. Other types of traps are not known here or in the surrounding region. The Sundance, Cloverly, and "Muddy" sandstone member of the Thermopolis are the main potential producing horizons. A test well to the pre-Cambrian in Big Medicine Bow anticline (Pl. 2) about 10 miles east of the northern part of the district showed no oil in the Tensleep (Shoenfelt, 1938, p. 35). Small yields have been obtained from the Wall Creek sandstone in Rock Creek anticline located 20 miles east of Elk Mountain (Dobbin, *et al.*, 1929b, p. 147).

Bloody Lake anticline.—Dips of the Mesaverde on the flanks and nose of the anticline and in the northern part of Fort Halleck syncline seem to indicate a structural high in sec. 4. The fault on the west flank of the anticline, which has been interpreted as formed by tension fracturing followed by propagation of the fracture downward nearly parallel to bedding, indicates that similar faults may occur in the area partially enclosed by the Mesaverde. Probably the Steele is extensively drag folded, in which case structural detail near the surface does not necessarily indicate structure at potential producing horizons. The Kasoming well cannot be considered as an adequate test. The generalized driller's log shows that the well was abandoned at 4225 feet without reaching the "Muddy" or Cloverly. Failure to encounter one of these can be explained by location of the well off structure, or by deflection of the drill down one of the flanks, or by thickening of shales on the crest of the anticline.

Simpson Ridge and Pass Creek anticlines.—Strikes and dips of the Shannon sandstone in secs. 13, 14, and 24, T. 20 N., R. 81 W., indicate a dome, locally known as the Mill Creek structure, in the western part of sec. 13 and the eastern part of sec. 14. Judging from the writer's map (Pl. 1), the Midwest well in sec. 11 is north of the high point. The well penetrated 1300 feet of Niobrara and yielded oil-saturated cores from the "Muddy" or Cloverly. The double thickness of Niobrara indicates drag folding or faulting near the crest of the anticline. The Steele beneath the alluvium is probably affected by numerous folds and faults formed by crumpling of the shales a few hundred feet below the competent Mesaverde sandstones. Such folds and faults would not necessarily be present in the more competent beds in depth. The occurrence of a bedding thrust in the Mowry shale of the west flank of Pass Creek anticline, as shown in sections J-J′ and K-K′, indicates the possibility of a similar thrust at the same horizon in the Mill Creek structure. Such a thrust should die out parallel to stratification west of the anticline and in a flexure east of the crest of the anticline, possibly without reaching the present land surface. It should offset the axial





CONVENTIONAL SYMBOLS

MAP

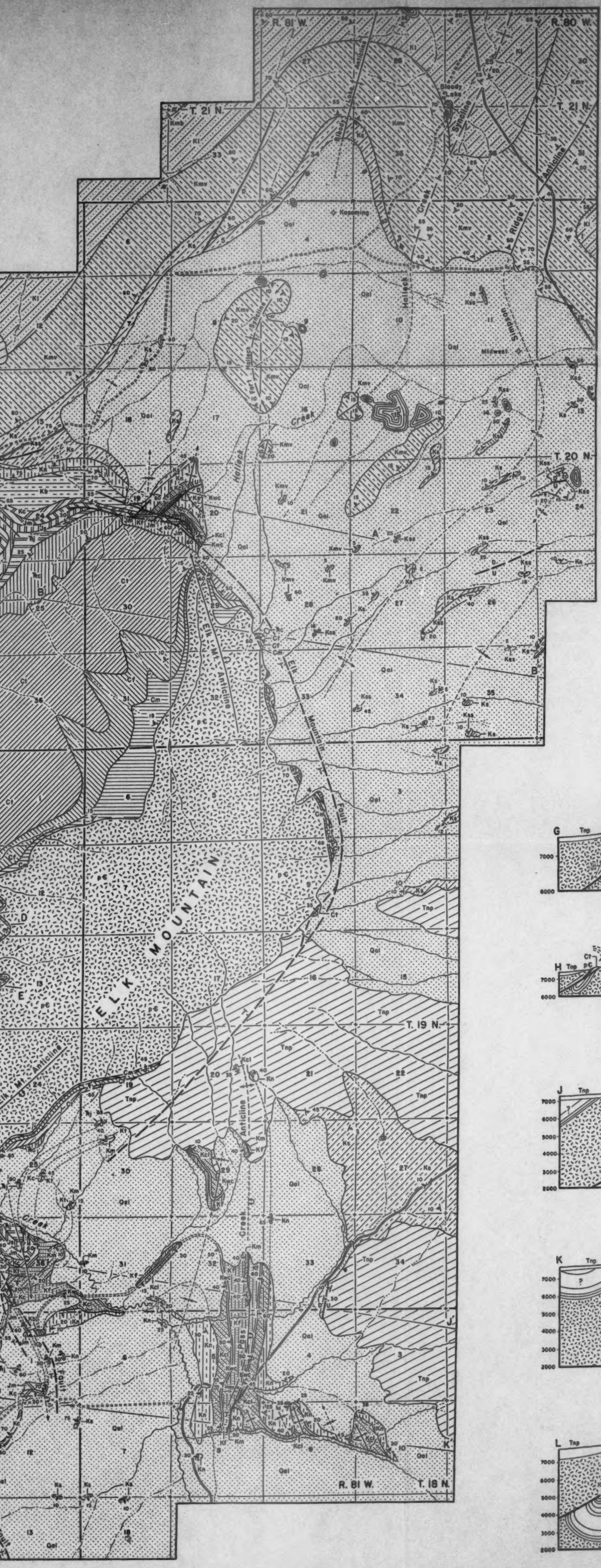
- Paved or graded road
- Secondary road
- Permanent stream
- Intermittent or ephemeral stream
- Lake
- Marked section corner
- Formation boundary
- Formation boundary covered by later deposits
- Strike and dip of beds
- Strike and dip of overturned beds
- Strike of vertical beds
- Axis of anticline
- Axis of syncline
- Axis of overturned anticline, showing direction of dip of axial plane
- Axis of fold covered by later deposits
- Direction of pitch of fold
- Known fault
- Known fault covered by later deposits
- Dip of fault
- Thrust fault; T is on block above fault
- U, upthrown, and D, downthrown, blocks along high-angle fault
- Relative direction of movement parallel to strike of high-angle fault

SECTIONS

- High-angle fault. Arrows show relative direction of movement in plane of section. A, relative movement away from observer; T, toward observer
- Thrust fault

SCALE

0 1 2 3 Miles


R. 81 W.
R. 80 W.
T. 21 N.
T. 20 N.
T. 19 N.
T. 18 N.
ELK MOUNTAIN
Elk Mt. Anticline
Elk Mountain Fault
Hollack Creek
Slimpson Creek
Bloody Lake Anticline
Bloody Lake Syncline
Kaseming
Midwest
Qal
Kmv
Kss
Ks
Kl
Kmb
Kc
Kcl
Kwc
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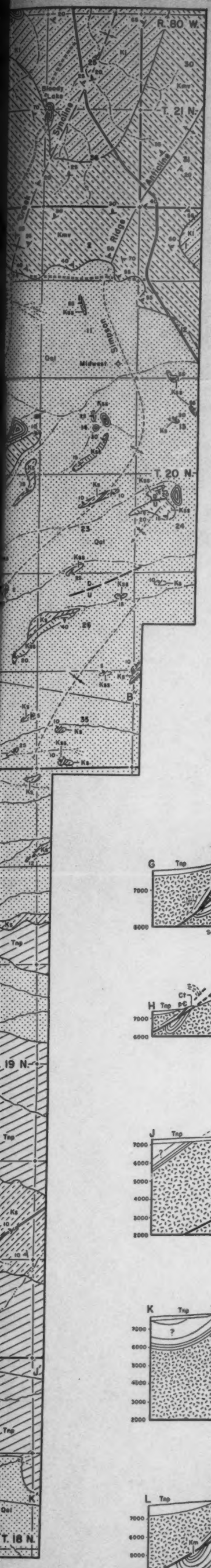
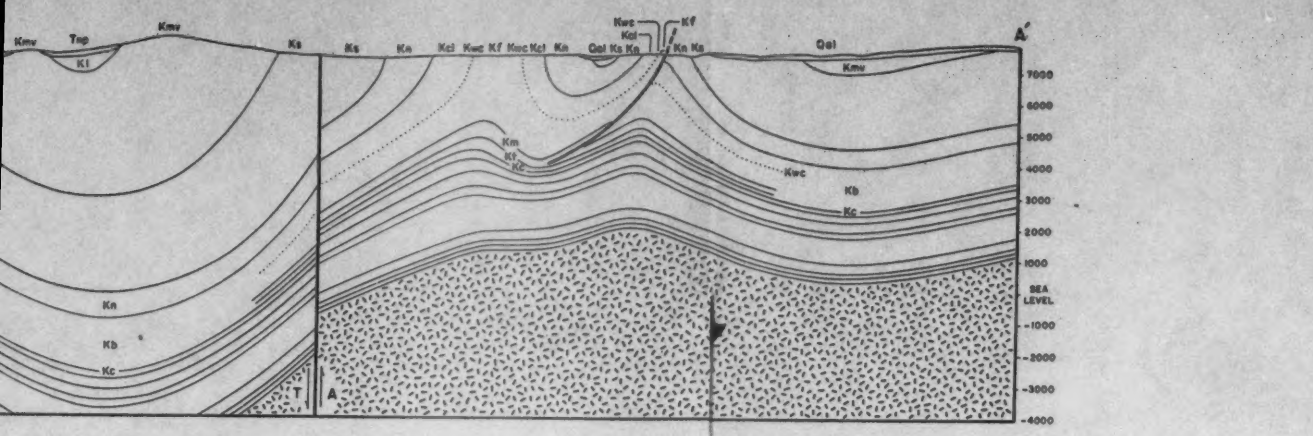
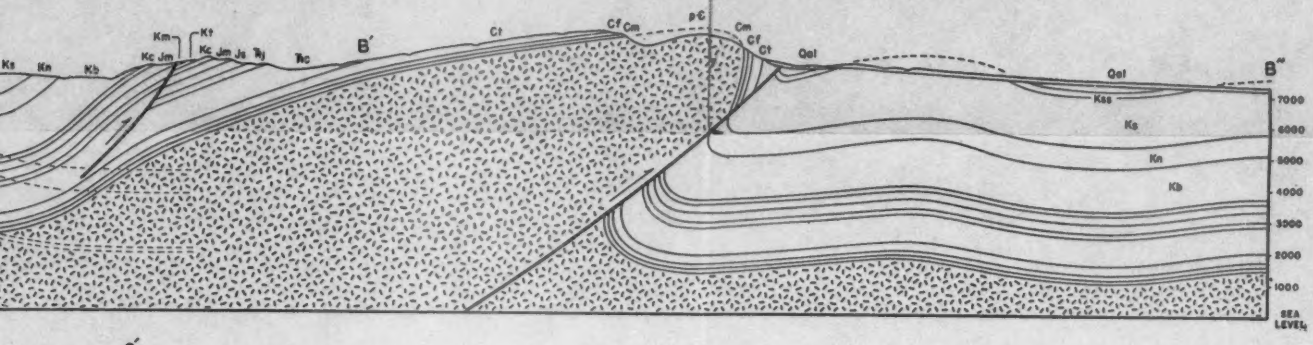
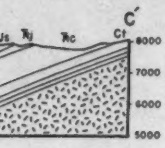
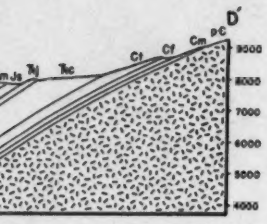


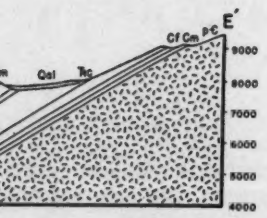
C MAP AND STRUCTURE SECTIONS OF THE ELK MOUNTAIN DISTRICT

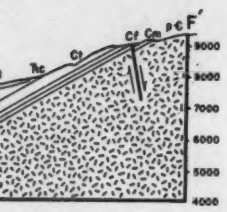


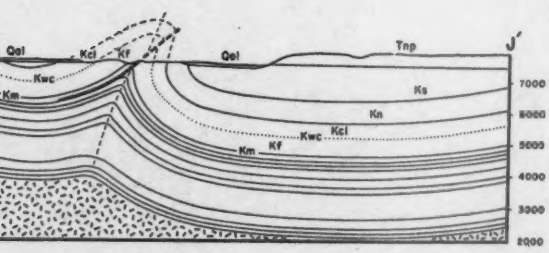


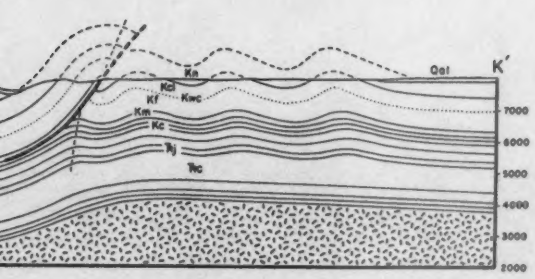


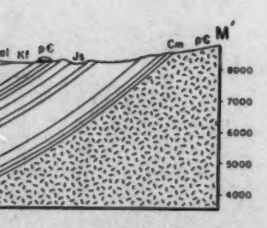












SEDIMENTARY ROCKS

	Qal	Quaternary alluvium	
MIOCENE (?)	Tnp	North Park formation	
	Kmb	Medicine Bow formation	
	Kl	Lewis shale	
	Kmv	Mesaverde formation	
	Ks	Steele shale	Kss Shannon sandstone member of Steele
	Kn	Niobrara formation	
UPPER CRETACEOUS	Kcl	Carlile shale	
	Kwc	Wall Creek sandstone	
	Kf	Frontier shale	Kb Benton group
	Km	Mowry shale	
	Kk	Thermopolis shale	
LOWER CRETACEOUS	Kc	Cloverly formation	
UPPER JURASSIC	Jm	Morrison formation	
	Js	Sundance formation	
TRIASSIC	Tj	Jelm formation	
PERMIAN	Tkc	Chugwater formation	
PENNSYLVANIAN	Cf	Tensleep sandstone	
	Cf	Fountain formation	
MISSISSIPPIAN	Cm	Madison limestone	

IGNEOUS AND METAMORPHIC ROCKS

PRE-CAMBRIAN	p€	Granites, gneisses, and schists

plane so that the part of it above the thrust is east of the part below. A vertical boring at the axial plane just above the thrust would continue downward into steeply dipping younger beds of the east flank. The double thickness of Niobrara in the Midwest well is consistent with the occurrence of a thrust of this type.

From sec. 33, T. 20 N., R. 81 W., to sec. 23 Simpson Ridge anticline pitches northeast. The thick cover of gravel and boulders at the base of Elk Mountain conceals relations of the fold to the Elk Mountain block. The beds of the east block may change strike abruptly and be truncated by the thrust. If this is so, the thrust fault provides closure at the southwest end of the anticline, and the high points of beds in the anticline are beneath the pre-Cambrian. A well northwest of the surface trace of the axial plane close to the pre-Cambrian should reach the crest of the fold at a potential producing horizon considerably below its high point.

Another possibility is that Simpson Ridge anticline and Pass Creek anticline are parts of a continuous fold extending beneath Elk Mountain thrust, in which case the Elk Mountain block must have moved across a syncline. This relation does not seem improbable because the pre-Cambrian of the northeastern part of Coad Mountain lies on the sediments of the west flank of an anticline near the west boundary of sec. 36, T. 19 N., R. 82 W., and line J-J′. Section L-L′ also shows thrusting of an anticline across an adjacent syncline. Another possible indication of continuity of the two anticlines is the northward pitch shown by the northernmost exposures on Pass Creek anticline and the southernmost on Simpson Ridge anticline. If the folds are continuous and the pitch is constantly to the north beneath the thrust, the structural high is on Pass Creek anticline. The "Muddy," Cloverly, and Sundance are here so close to the surface that there is considerable possibility of escape of oil and gas, particularly in view of the occurrence of bedding thrusts in the Mowry at a number of places on the west flank of the anticline. The thrusts cut the fold so that in places there is no axial plane at the surface and the axial plane beneath the thrust is offset west from its position above the thrust.

CONCLUSIONS ON ELK MOUNTAIN DISTRICT

The general northerly trends of folds and strikes of thrusts in the Elk Mountain district indicate that deformation was caused by east-west compression with predominant upward relief. In detail, however, the structural pattern differs markedly from that produced experimentally by Mead (1920), who coated a stretched rubber sheet with paraffin or

wax and permitted the sheet to contact. Under these conditions folds trend and thrust faults strike normal to compression; thrusts die out laterally in the limb between anticline and syncline. Many structural features in the Elk Mountain district, such as thrusts that pass laterally northward into west-striking tear faults, Bear Butte thrust cutting diagonally across a major syncline and anticline, and the northeast-striking Rattlesnake Pass fault, strongly suggest features that developed near the laterally expanding free edges of Mead's sheets and also those produced in a pressure box by Link (1928) under conditions of nonuniform compression applied to a homogeneous succession of artificial sediments, or uniform compression applied to a laterally varying succession. In the Elk Mountain district there is little facies variation in the sedimentary rocks. The pre-Cambrian igneous and metamorphic rocks here and in the surrounding region change markedly in character from place to place. The writer believes, therefore, that the structural pattern of the Elk Mountain district was produced by east-west compression applied to a heterogeneous basement complex and a relatively thin cover of laterally uniform sedimentary rocks. The pattern may have been influenced by northward lateral relief afforded by sinking of the basin to the north under the load of sediments derived from erosion of the rising mountains.

RELATION TO STRUCTURE OF SURROUNDING REGION

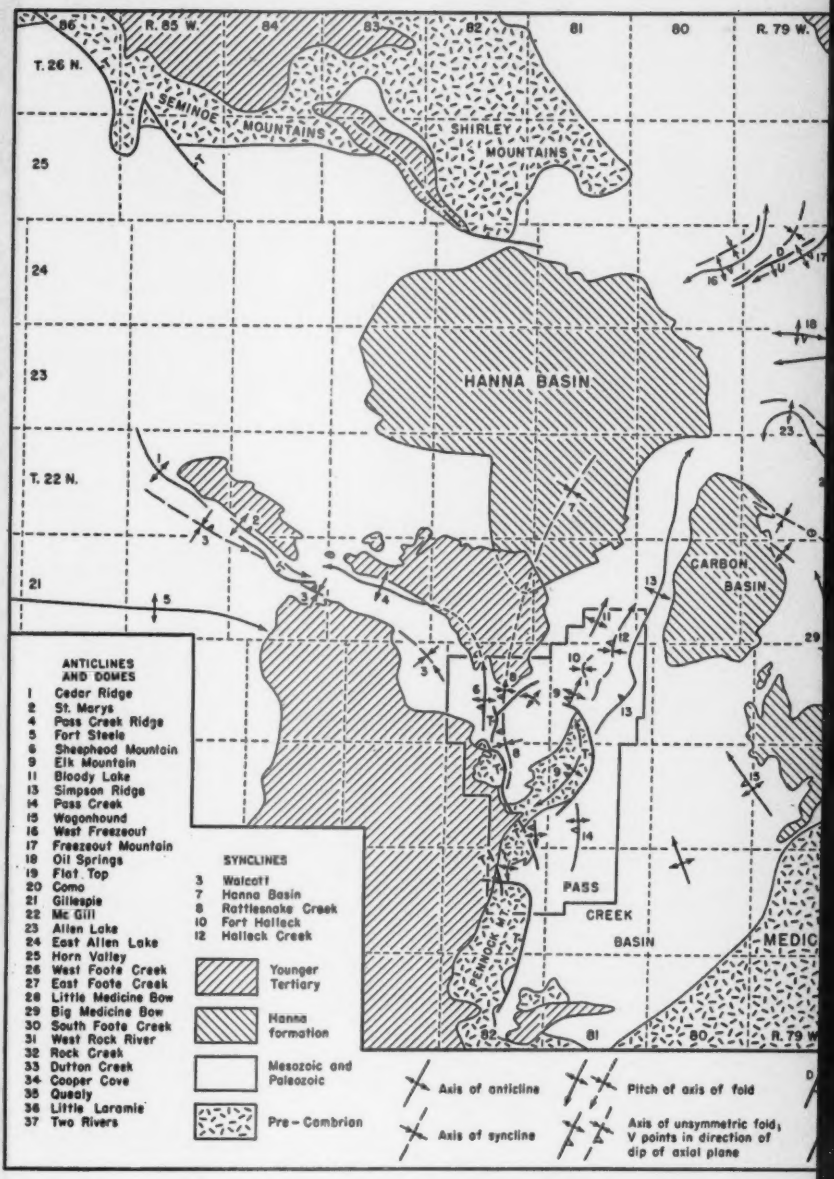
Description.—The map (Pl. 2) compiled from sources marked by an asterisk in the list of references shows the following:

(1) Folds in the Elk Mountain district trend north. Axial planes of unsymmetric folds and thrust faults dip west. Folds radiate in a fanlike arrangement from the northern part of the district.

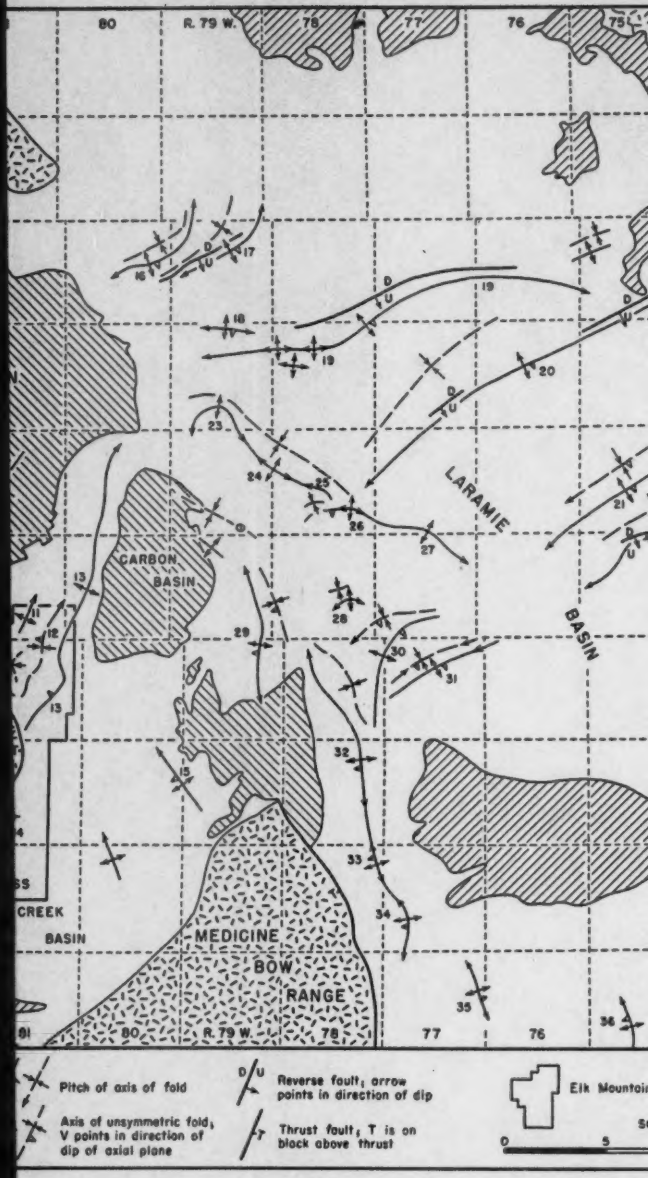
(2) Thrust faults along the southern margin of the Seminoe and Shirley Mountains, which form the eastern part of the Sweetwater Arch (Fig. 1), dip north or northeast.

(3) Folds radiating from the western part of the Elk Mountain district trend west or northwest; axial planes of unsymmetric folds and a thrust fault dip northeast. Structures here suggest a genetic relation with the deformation processes that affected the southern margin of the Seminoe and Shirley Mountains.

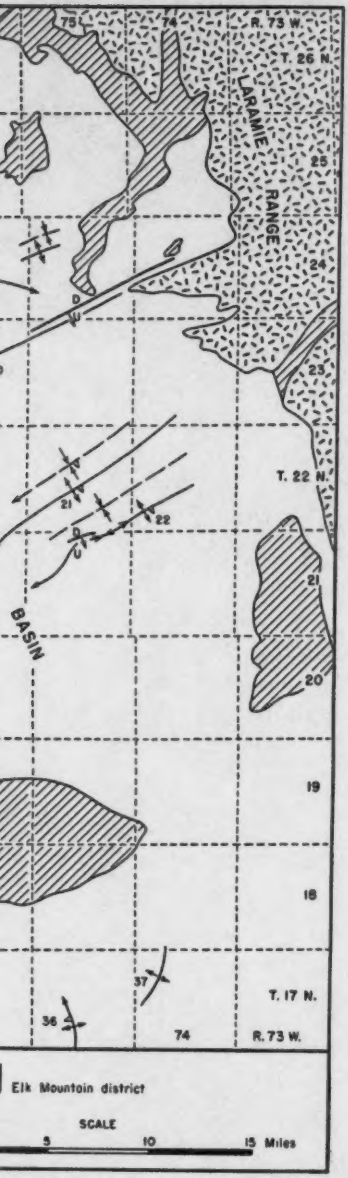
(4) East of the Medicine Bow Range folds trend north. The axial planes of most of them dip west; the axial plane of only one, Quealy dome in T. 17 N., R. 77 W., dips east. The Arlington thrust fault, at the east base of the Medicine Bow Range, dips west. Folds extending north from the eastern border of the range swing to the northwest, and this trend is markedly shown in the northern part of the Carbon Basin and the area a few miles northeast.



TECTONIC MAP OF THE REGION AROUND THE NOR



ON AROUND THE NORTH END OF MEDICINE BOW RANGE


75
74
R. 73 W.
T. 26 N.
LARAMIE RANGE
25
24
23
T. 22 N.
21
22
BASIN
21
20
19
18
37
T. 17 N.
36
74
R. 73 W.
Elk Mountain district
SCALE
5
10
15 Miles

RANGE

(5) Folds in the northeastern part of the Laramie Basin trend northeast, and their axial planes dip southeast. The northwest limbs of several anticlines are cut longitudinally by reverse faults or high-angle thrusts dipping southeast. This type of structure extends into T. 24 N., Rs. 79 and 80 W., within 12 miles of the northward-dipping thrust at the base of the Shirley Mountains. Surface traces of axial planes of folds are here convex to the southeast, yet a reverse fault dips in this direction. In the western part of the Laramie Basin in Ts. 20 and 21 N., Rs. 77 and 78 W., folds with northeast trend and southeast dip of axial plane occur only a few miles east of folds with north to northwest trend. Another fact not shown by the map is that the northeast-trending folds affect only the lower Mesaverde and older rocks in the lower 7000 feet of the sedimentary succession.

Interpretation.—The southern part of the Rocky Mountains in Colorado and New Mexico trends nearly north, and the northern part in Montana and Alberta fairly consistently trends northwest. In the Central Rockies of Wyoming, northwestern Colorado, and northeastern Utah major ranges extend west, northwest, and north. Minor folds have the same trends, and in addition a few trend northeast. Such an irregular arrangement cannot be explained by uniform compression or a simple shear couple such as that postulated by Chamberlin (1940) to account for trends of ranges and minor folds in northern Wyoming and southern Montana. It seems best to consider structural relations around the north end of the Medicine Bow Range in terms of effective compression, either primary or resolved in a force couple, and relative directions of movement of major mountain blocks. The following hypothesis is advanced to account for relations shown on the map (Pl. 2).

The predominant stress in the southern part was nearly east-west compression such as has been inferred for the Southern Rocky Mountains. If one accepts the hypothesis that thrust faults and axial planes of folds dip toward the active force, pressure was from the west. The gradual change of trend in the belt extending from the east border of the Medicine Bow Range to T. 22 N., Rs. 78 and 79 W., strongly suggests marginal folds wrapping around the north end of a relatively eastward-moving resistant mass.

Effective compression west of the Hanna Basin was close to north-south. Localization of sharp folding and thrusting in the belt southwest of the Hanna Basin is genetically connected with relative southward movement of the Rawlins uplift (Fig. 1), in which the pre-Cambrian core of the Fort Steele anticline appears at the surface.

In the Medicine Bow Range a few miles south of the Pass Creek Basin, bedding, schistosity, and a major fault in a thick succession of metamorphosed sediments and extrusives strike northeast and dip steeply to the southeast (Blackwelder, 1926). In the northern part of the Laramie Range foliation in schists strikes northeast, and contacts between schists and intrusive rocks are as a rule steeply inclined or vertical (Spencer, 1916, p. 52). It is believed that the folds trending northeast in the Laramie Basin were caused by Laramide deformation of weaker rock belts in the pre-Cambrian basement complex. North-south compression initiated vertical movement on planes of weakness dipping southeast. Relative upward movement of the hanging walls of the faults thus formed produced northwest-southeast shortening. Some of the faults were propagated far enough up into the sediments to reach the present land surface. These are confined to the northeastern part of the Laramie Basin, from which beds younger than lower Steele have been eroded. In Ts. 20 and 21 N., Rs. 77 and 78 W., where Steele and younger rocks are exposed, faults do not reach the surface. The folds near the east end of the Shirley Mountains suggest marginal folds changing trend around the east end of a relatively southward-moving mass, in which case faults and axial planes of folds should dip northwest. Dips are, however, to the southeast toward the outside of the arc, indicating that control was exercised by the structures in the pre-Cambrian.

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ORDOVICIAN STRATIGRAPHY OF CENTRAL VERMONT

BY LOUIS W. CURRIER AND RICHARD H. JAHNS

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ABSTRACT

The discovery of a crinoidal limestone near Northfield has necessitated a drastic revision of the stratigraphy in central Vermont east of the Green Mountain axis. The "Cambrian-Ordovician" boundary of earlier papers proves to be a horizon probably within the Ordovician system; it marks the base of the Northfield slate, which is tentatively correlated with the basal beds of the Tomifobia formation (Middle Ordovician?) in Quebec. Two new formations are defined for central Vermont—the Shaw Mountain formation, which contains the crinoidal limestone and lies immediately beneath the Northfield slate, and the Cram Hill formation, which is tentatively correlated with the upper part of the Magog slates of Quebec and thus is probably of Middle Ordovician age. The Northfield slate of Richardson is redefined with respect to the "Memphremagog slate" of earlier usage.

The earlier interpretation of Beekmantown age for the Northfield slate is held untenable because of its stratigraphic relations to the Shaw Mountain formation and because the supposed graptolites upon which the earlier correlation was made have now proved to be inorganic features.

The Irasburg conglomerate is an intraformational bed within the Waits River formation (Middle Ordovician or later?) and as such has no important regional significance.

INTRODUCTION

The fossiliferous Paleozoic rocks in Vermont west of the Green Mountains have drawn the attention of geologists for a century or more, and the published results of investigations constitute a notable portion of the geologic literature of New England. The great assemblage of metamorphosed sedimentary rocks in central Vermont east of the Green Mountains, on the other hand, received comparatively little attention during the nineteenth century, and, indeed, very little detailed mapping was done in the region before 1920. This was apparently due to the almost complete lack of recognizable fossils, the greater complexity of the structure, and the relatively high degree of metamorphism, all of which made it particularly difficult to separate formations and correlate them with reasonable accuracy. It has long been felt that many of the rocks east of the Green Mountains axis are chronologic equivalents of the better known fossiliferous rocks to the west, but the definite correlation of the beds in the two areas has not been made, largely because of the difficulties just mentioned.

Edward Hitchcock's early monograph (Hitchcock, E., and others, 1861) on the geology of Vermont stands today as the most complete treatment of these complex rocks. An extensive series of reports somewhat more detailed than Hitchcock's was published later by C. H. Richardson,[1] who traversed a group of townships from Canada southward to the Massachusetts State line. These townships lie along the trace of his so-called "Cambrian-Ordovician erosional unconformity," which is in reality a lithologic boundary between argillaceous and arenaceous rocks on the west and dominantly calcareous and argillaceous types on the east (Fig. 1). More general reports by Richardson (1902; 1906; 1919a; 1929a), and several

[1] *See* papers on areal geology in annual reports of the Vermont State Geologist, 1908-1938.

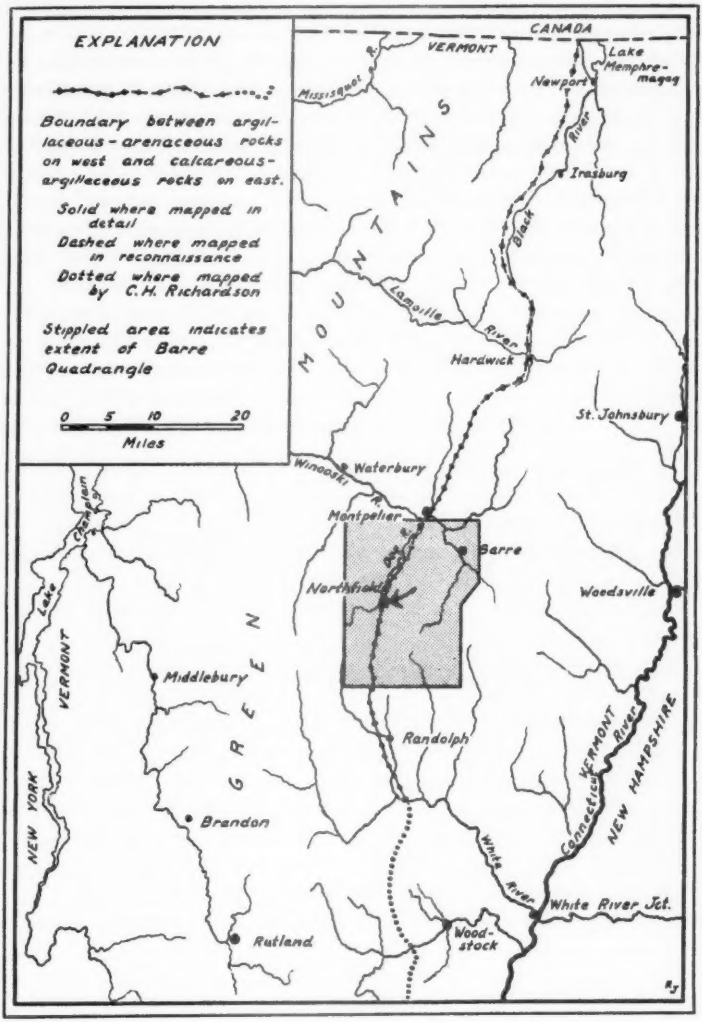

FIGURE 1.—*Index map of central and northern Vermont*

Showing location of the Barre quadrangle. Arrow indicates location of area shown in Figure 4.

recent papers by Foye (1919), Gordon (1927), Perry (1929), and Hawkes (1941) that deal with areas east and northeast of Rutland, complete the list of significant contributions to the literature on central Vermont.

Most previous investigators have agreed on the broader questions of stratigraphy and structure of the rocks east of the Green Mountains, but many important details are still confused. Furthermore, until recently there has been a conspicuous lack of specific data that should form the basis of proof for the many general conclusions advanced. The result has been an increasing tendency among thoughtful geologists to question major structural and stratigraphic hypotheses that were regarded as more or less factual a decade ago and to demand more detailed information for many critical areas before accepting the generalizations. It is the purpose of this paper to supply basic data for a relatively small part of the stratigraphic column in central Vermont that contains recognizable marker beds and, by means of these data, to establish a firmer foundation for future studies.

FIELD WORK AND ACKNOWLEDGMENTS

The information contained in this paper is based on widespread reconnaissance observations over several quadrangles made by Currier during field seasons from 1934 to 1938 and on detailed mapping by Jahns of an area of approximately 250 square miles in the Barre and adjacent quadrangles (Fig. 1) during parts of the seasons 1937 to 1940 inclusive. The "Cambrian-Ordovician" contact as mapped by C. H. Richardson extends through this area.

Currier's part of the investigation in the period noted above was financed by a grant from The Geological Society of America for a general study of the metamorphic rocks. Recognition of the necessity for more detailed structural and stratigraphic data led to a project for mapping the Barre quadrangle, and this part of the work was carried on by the U. S. Geological Survey and under Currier's general supervision. Jahns was originally assigned to this project as field assistant but subsequently he took over all the detailed stratigraphic mapping. The authors have conferred in both the field and office, but the separation of the various stratigraphic units is based almost entirely on Jahns' field work.

STRATIGRAPHY

GENERAL STATEMENT

The sedimentary formations discussed in this paper represent a small part of a very thick series of Paleozoic strata. Certain of these formations have distinctive lithologic characteristics that make it possible to trace them for long distances along the strike of the beds. The formations dis-

cussed include a transition zone between dominantly calcareous strata to the east and almost entirely noncalcareous strata to the west. It is in this transition zone that the so-called Cambrian-Ordovician contact of Richardson lies. The contact was interpreted by him as an erosional unconformity. Recent discovery of fossils within the transition beds has necessitated a revision of his earlier interpretations. This paper deals particularly with the stratigraphic relations, significance, and implications of this fossiliferous horizon.

Three main stratigraphic units have been recognized in the belt that includes the transition beds and the "Cambrian-Ordovician contact" of Richardson. These units are the Cram Hill formation, the Shaw Mountain formation, and the Northfield slate (Fig. 2). The Cram Hill formation is composed chiefly of greenish-gray phyllite, greenish siliceous schist, intercalated thin beds of quartzite, and clastic volcanic rocks. The Shaw Mountain formation consists of a basal quartz conglomerate, a rhyolite tuff, and a thinly bedded crinoidal limestone. The Cram Hill and Shaw Mountain formations are recognized and defined here for the first time. The Northfield slate, recognized and named by Richardson, is here redefined with respect to the "Memphremagog slate" (Richardson) and the Magog slates of Quebec.

The Northfield slate is overlain by the very thick Waits River formation, largely the Waits River limestone of earlier reports, but here termed a formation because of the prominent beds of phyllite and quartzite included in it.

Beneath the Cram Hill formation lies a very thick series of arenaceous schists, quartzites, phyllites, and slates, at least some of which are probably of Cambrian age.

All these sedimentary rocks have been greatly deformed and metamorphosed. With no important exceptions, their degree of metamorphism places them in the low-grade and middle-grade zones as defined by Tilley (1925) and Harker (1939) and applied in New Hampshire by Billings (1937, p. 470, 590, *et seq.*). In the Barre quadrangle and adjacent areas the intensity of metamorphism increases gradually and progressively from the north and west toward the south and east. These changes seem not to depend upon the distribution of rocks of different ages but are rather the result of metamorphic processes that acted after the deposition of the youngest sedimentary rocks in the area.

Although the entire stratigraphic section has been deformed so that the beds are now vertical or dip very steeply, the major structure appears to be simple. The rocks occur as a thick, homoclinal series, which presumably represents the east flank of the great Green Mountain arch. Isoclinal folds are exceedingly common but with few exceptions are very small; the

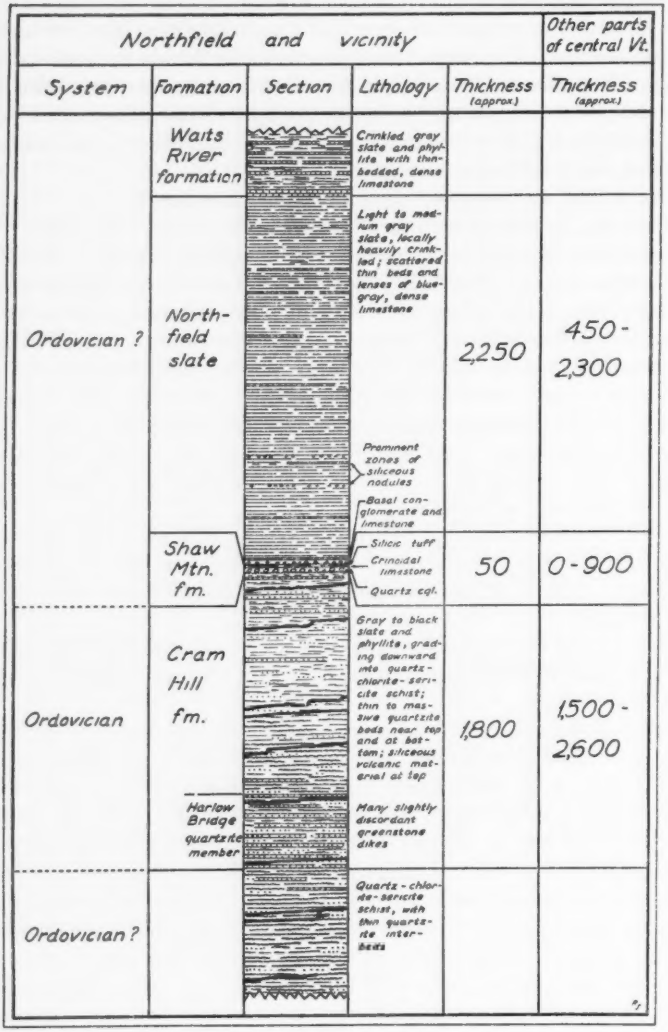

FIGURE 2.—*General columnar section for rocks in the vicinity of Northfield, Vermont*

absence of larger folds and the consistently steep dip of the beds are responsible for the distribution of the various stratigraphic units as rather straight, subparallel bands whose traces are affected little or not at all by topography. The general strike of these beds is north-northeast between Northfield and the Lake Memphremagog region (Fig. 1) except in the vicinity of Hardwick, where they trend broadly eastward. South from Northfield the strike shifts gradually to north-northwest in the Barre quadrangle but still farther south it is again north-northeast.

CRAM HILL FORMATION

Lithology.—The Cram Hill formation consists of several rock types which grade into one another both across and along the strike. This is believed to be a feature of original deposition rather than of later deformation. The most characteristic rock type within it is a splintery, greenish-gray to black phyllite, which weathers gray to brown, and it is composed chiefly of quartz, sericite, and chlorite, with appreciable accessory carbon and pyrite. Locally it grades into a gray to black slate. (Pl. 1, fig. 1.) It is typically exposed along the Bull Run road south of Northfield (Barre quadrangle). This rock also forms the falls of the Dog River at Northfield and Northfield Falls.

Distinctive platy to massive quartzite beds mark the base of the Cram Hill formation. This basal member is here termed the Harlow Bridge quartzite because of its excellent exposures in the vicinity of a large railroad bridge of that name 2.4 miles south-southwest of Northfield. It has been mapped along the entire traced outcrop of the Cram Hill formation. It maintains a remarkably constant thickness north of the type locality but thickens appreciably to the south.

The thin beds of quartzite near the top of the section are not mappable units. The Harlow Bridge quartzite, however, is a distinctive member and is the sole means of discrimination between the arenaceous green schists of the Cram Hill formation and those of older formations to the west. This member consists of buff to pale-green, massive to thin-bedded quartzite with intercalated siliceous schist.

Pale to grassy-green quartz-sericite-chlorite schist is dominant in the Cram Hill formation north of Northfield Falls, where phyllitic rocks appear only in the uppermost parts of the section. The schist is very siliceous, and thin, platy, greenish quartzite beds are common within it (Pl. 1, fig. 2). Rocks of this type are well exposed along State Highway 12 north of Riverton. A very distinctive, silvery and greenish sericite phyllite is prominent in the lower part of the section between Harlow Bridge and Riverton; it contains abundant white mica and in the past has been mistaken for a talcose rock because of its softness.

Northward along the strike this rock passes into the typical green siliceous schist; southward it grades into dark gray phyllite.

The Cram Hill formation also includes clastic volcanic rocks and thin greenstone dikes. The volcanic rocks are flows, breccias, and tuffs and occur only in the upper part of the formation. In general, they are poorly exposed. The principal silicic volcanics include soda rhyolite that contains conspicuous crystals of quartz and albite, a highly siliceous breccia or conglomerate (that may be in part water-laid), and a platy rhyolitic rock that contains bluish milky quartz crystals. These rocks are gray to buff where freshly exposed but in most places have weathered deeply and are light brown. Very few undoubted flow rocks have been seen. Other volcanics are of basic types and probably were originally of basaltic or andesitic composition but now appear as dense and agglomeratic greenstones. They are greatly altered, massive, or schistose rocks composed mainly of secondary minerals—albite, chlorite, epidote, dolomite, and ankerite. Some of the basic volcanics appear to be flows, though others may be tuffs. The best exposures of these volcanics were found in the banks of the Dog River, at the railroad bridge 2½ miles southwest of Montpelier, in the Barre quadrangle.

Greenstone dikes are locally abundant. They are clearly younger than the volcanics for they cut the Cram Hill and older formations, including the volcanic beds, at sharp angles. The dikes consist chiefly of secondary minerals—epidote, chlorite, and ankerite—and are strikingly similar to the dikes found in the Albee formation of New Hampshire (Billings, 1937, p. 512-515). Excellent exposures of these dikes were found in fresh cuts along State Highway 12, north of Riverton.

A generalized section of the Cram Hill formation as it appears in the vicinity of Northfield is shown diagrammatically in Figure 2. Its lithologic character as shown is similar to that at the less easily accessible type locality, the east slope of Cram Hill 3½ miles south-southeast of Roxbury (Barre quadrangle). The interbedding of several different rock types is especially noteworthy, but a mineral assemblage characteristic of low-grade metamorphism—chlorite, sericite, epidote, albite, and dolomite—is characteristic of all types north of Braintree Hill in the Randolph quadrangle. Hence they correspond to the chlorite zone of Tilley (1925) and Harker (1939). Biotite porphyroblasts are found in the slaty and phyllitic rocks on Braintree Hill, however, and farther south, near Randolph, the transition into middle-grade metamorphism is further indicated by garnet porphyroblasts in the dark phyllites, garnet and actinolite crystals in the green schists, and actinolite and oligoclase-albite in the greenstones. The biotite is present in considerable quantity. Thus within a distance of about 3 miles along the strike, rocks of the Cram

FIGURE 1. SLATY FACIES OF CRAM HILL FORMATION
1.8 miles south-southwest of Northfield. Bedding and slaty cleavage are parallel.

FIGURE 2. QUARTZ-SERICITE-CHLORITE SCHIST FACIES OF CRAM HILL FORMATION
2.7 miles west-southwest of Montpelier. View looking north. Bedding is essentially vertical, foliation dips steeply west. These rocks represent a stratigraphic horizon approximately equal to that of slaty facies of Figure 1.

FACIES OF CRAM HILL FORMATION

IGURE 1. RIVER-SCOURED SURFACE OF ONGLOMERATE AT BASE OF NORTHFIELD SLATE
ank of Dog River 2.5 miles southwest of ontpelier. Note elongation of fragments. View looking down and northward.

IGURE 2. BASAL QUARTZ CONGLOMERATE OF SHAW MOUNTAIN FORMATION
illside east of Northfield Falls. Note pro- unced stretching of pebbles. View looking eastward at a bedding plane.

ORDOVICIAN CONGLOMERATE BEDS IN CENTRAL VERMONT

Hill formation pass from the chlorite zone through the biotite zone into the garnet zone, to use the terminology of Tilley (1925) and Harker (1939), or from the low- to middle-grade zone of metamorphism as used by Billings (1937).

Thickness.—The strata of the Cram Hill formation are vertical or nearly so, and it would therefore appear that the width of outcrop should be a close measure of the true thickness of the formation. This is not so, however, because the beds have been deformed by intricate minor folding and faulting. Most of the outcrops reveal this deformation by many small isoclinal folds that commonly are sheared off on either or both sides; hence there has been a considerable apparent thickening of the strata by small-scale folding and by shearing movements nearly parallel to the strike. The great abundance of small and minute folds and shears and the lack of continuous large outcrops preclude, therefore, any accurate determination of the thickness.

The approximate thickness was determined, however, by extrapolating data obtained from many outcrops. Careful sketches were made of the folds and shears in 96 exposures in which the structures could be traced by following individual and clearly recognizable beds. An average factor of repetition was obtained from these sketches and also from direct measurements on exposures, and the two averages showed a remarkable consistency in the ratio of breadth of outcrop to thickness of the beds. For this reason, the average ratio so determined has been applied to the entire formation. The breadth of outcrop multiplied by 0.4 to 0.5 is considered to be the approximate thickness of the Cram Hill formation.

Between Northfield and Randolph the formation is from 1500 to 2600 feet thick, probably with a general average of about 1700 feet. North of Northfield it thickens progressively, but this thickening is attributed chiefly to the presence of countless greenstone dikes. The Harlow Bridge quartzite member, on the other hand, is less than 300 feet thick 2 miles northeast of Montpelier Junction but thickens southward at a rather uniform rate, and on the east slope of Belcher Hill, 8.5 miles south-southwest of Northfield, it is more than 800 feet thick. Inasmuch as this thickening appears to be due almost entirely to original deposition and is accompanied by a corresponding thinning of the overlying finer-grained strata, the base of the Harlow Bridge member, and thus of the Cram Hill formation itself, may be a reasonably good time boundary.

Distribution.—The Cram Hill formation has been traced from Randolph to a point 2½ miles north of Montpelier, or for a total distance of approximately 27 miles. It is well exposed on the east slope of Cram Hill 3½ miles south-southeast of Roxbury (Barre quadrangle) and re-

ceives its name from this locality. The formation is crossed by the Winooski River between Montpelier and Montpelier Junction (Montpelier quadrangle) and follows the Dog River and Bull Run from Montpelier Junction southward to Cram Hill. South of this point it forms prominent highland areas, including Ferry Hill and a part of Braintree Hill (Randolph quadrange).

Age.—No fossils have been found in the Cram Hill formation in central Vermont. The rocks of this formation constitute the upper part of Richardson's "Missisquoi schists" (Richardson, 1919b, p. 137-138) which at first he tentatively placed in the Lower Cambrian but later (Richardson, 1927) interpreted as Upper Cambrian. This is the most recent age assignment made hitherto and was based mainly upon the presence of the "Missisquoi schists" (later termed "Missisquoi group") unconformably below the "Memphremagog slates" (Richardson, 1916, p. 142-145; Richardson and Camp, 1919), which were believed to be of lower Trenton to Deepkill (Beekmantown) age.

The writers believe that the Cram Hill rocks, and consequently the upper portion of the "Missisquoi group," are not Cambrain but are probably Middle Ordovician and that a lower Trenton age for the "Memphremagog slates" of Richardson is open to question. The specific reasons for these new views are discussed in the sections on the Shaw Mountain formation, the Northfield slate, and Regional Correlations. The present assignment of the Cram Hill formation to the Ordovician system is based chiefly upon its correlation with a part of the fossiliferous Magog slates (of Trenton age) in southern Quebec, as well as upon its stratigraphic position (in central Vermont) immediately below beds containing a fauna that appears to be Middle Ordovician or younger.

SHAW MOUNTAIN FORMATION

General statement.—The name Shaw Mountain formation is suggested for a relatively thin group of beds that is well exposed near the base of the north slope of Shaw Mountain, 2 miles south-southwest of Northfield (Barre quadrangle). Because it is composed of distinctive lithologic units, at least one of which is fossiliferous, it has great stratigraphic importance.

Lithology.—The Shaw Mountain formation consists of three lithologic types (Fig. 3). These have been found at many places and at consistent stratigraphic horizons. The rock types are reasonably persistent along the strike, but exposures are generally small. Because of this discontinuity and the structural complexity of the beds it was necessary to make many close studies of the exposures before the stratigraphic rela-

tions could be definitely established. The relations so determined have been of great value in solving problems relating to the structure and sequence of adjacent formations.

The basal member is a resistant, massively bedded quartz conglomerate that crops out in high, rounded hillocks and prominent white to pinkish-brown cliffs. It is a unique type in central Vermont and therefore an exceptionally valuable key horizon. In several respects it is lithologically similar to the Clough conglomerate of New Hampshire, described by Billings (1937, p. 481-483), but is not necessarily of the same age. The pebbles are almost entirely of vein quartz, although after careful search greenstone, slate, phyllite, rhyolite, quartzite, and jasperoid have been recognized at several places. All pebbles have been drawn into elongated ellipsoids by strong tectonic action. The long axes of the ellipsoids range from ¼ inch to 11 inches in length and generally plunge very steeply to the north (Pl. 2, fig. 2). Locally, much larger boulders and pebbles constitute more than 90 per cent of the rock. At other places only a few scattered pebbles are present. The matrix is composed of quartz grains with subordinate flakes of sericite and chlorite and, in places, is very friable.

The following list includes the best and most representative exposures of the conglomerate in the Barre quadrangle, arranged in order from north to south:

West slope of Irish Hill (altitude approximately 1040 feet), 1.5 miles northeast of Northfield Falls.
Hillslope 0.6 mile east of Northfield Falls.
Hillslope 0.4 mile southeast of Northfield Falls. This is a large area with outcrops of intricately folded rocks.
Base of Turkey Hill 0.5 mile northeast of the Common in Northfield. This is an excellent example of a friable facies.
Lower slopes (altitude 800 to 1000 feet) on north side of Shaw Mountain, 0.6 mile east-southeast of Harlow Bridge school. (*See also* figure 4.)
Prominent knob at west end of curving ridge, 1.0 mile west of Bowman school. This knob and a high cliff down the slope immediately to the west are the most spectacular exposures of the conglomerate.

No exposures of the conglomerate were discovered north of Montpelier and south of Mt. Nevins (Barre quadrangle), so that known outcrops are restricted to a distance of approximately 18 miles along the strike.

The basal quartz conglomerate grades upward into a papery to platy soda rhyolite tuff composed chiefly of fine-grained quartz, alkali feldspar, sericite, and albite. Where fresh the rock is white to pale greenish white; in most places it has weathered very deeply to a light brown or tan, presumably owing to small amounts of altered magnetite and chlorite. The more sericitic beds weather to soft, pulpy masses and are poorly exposed. The more siliceous facies are tougher and more resistant, but their pronounced schistosity has prevented them from forming prominent ridges.

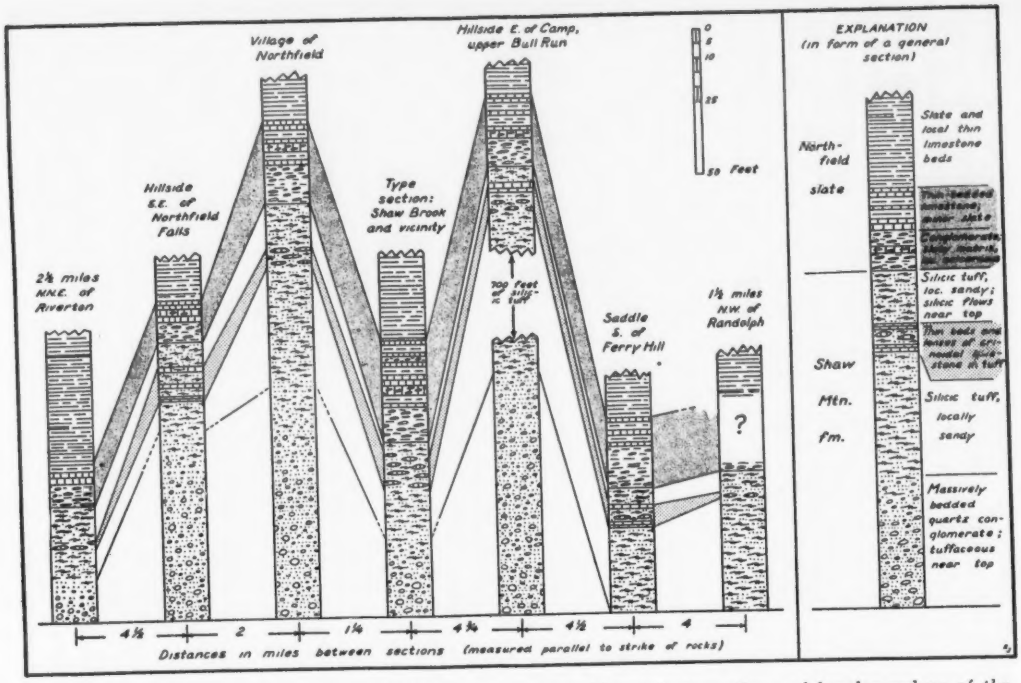

FIGURE 3.—*Variations in thickness and lithology of the Shaw Mountain formation and basal members of the Northfield slate*

The most important lithologic member is a white to slightly bluish crinoidal limestone that occurs as thin beds and lenses near the top of the formation. In most places the limestone is interbedded with tuff, but where the uppermost beds of tuff are absent the Northfield slate rests directly upon the limestone. The limestone is so thin and poorly exposed that many outcrops could be discovered only after its stratigraphic position was known exactly and careful search could be concentrated in small areas. A list of the most satisfactory limestone and tuff localities, from north to south, is as follows:

Small, west-flowing brook 1.0 mile west of crest of Irish Hill (limestone).
Hillslope 0.4 mile southeast of Northfield Falls. This includes the thickest known exposures of the limestone (limestone and tuff).
Brook 0.2 mile east-southeast of the Common in Northfield (tuff).
Shaw Brook 0.7 mile northeast of Harlow Bridge school (Fig. 4). This is the type locality for the limestone and for the crinoids it contains (limestone and tuff).
Sharp bend in road 0.45 mile east-northeast of Bull Run school (tuff).
Brook 0.2 mile northeast of camp on upper Bull Run. This brook exposes the thickest known section of the tuff.
Prominent exposure (tuff) near State Highway 12, 0.5 mile south-southeast of Old Church. This is near the mouth of Gilead Brook (Randolph quadrangle).

The northernmost exposures of the tuff were found in the vicinity of Riverton, where silicic flows and volcanic conglomerates appear at the tuff horizon and apparently occupy it entirely from this point north. To the south the tuff occurs at least as far as Bethel, and south of Randolph it contains small porphyroblasts of biotite. The crinoidal limestone has been found as far north as Montpelier and as far south as Randolph. It may extend farther in both directions but it has not yet been found beyond these limits despite careful search.

Thickness.—The thickness of the Shaw Mountain formation could not be determined by the method applied to the Cram Hill formation. The outcrop widths of the several members of the Shaw Mountain formation, especially of the quartz conglomerate, vary greatly within short distances along the strike; for example, a variation from 0 to nearly 200 feet in thickness of the conglomerate was found within a horizontal distance of 200 feet in several places. Because of these abrupt changes, as well as the abundant evidence of shearing in the beds, the irregular, pod-like distribution of the conglomerate is attributed to tectonic forces rather than to variations in original sedimentation. This interpretation is like that advanced by Billings (1937, p. 531-532) to explain similar features in the Clough conglomerate of New Hampshire. The quartz conglomerate of the Shaw Mountain formation is thought to have yielded to deformation as a very brittle mass and to have broken into separate blocks, whereas the incompetent tuffs flowed into the spaces between the blocks. Large-scale variations in thickness of the tuff, however, are thought to be

due in large part to original deposition, although it is not possible to separate the two factors entirely. Figure 3 represents the results of an attempt to measure the true thicknesses on the basis of the foregoing conclusions.

In general, the conglomerate thickens southward and probably reaches a maximum of 250 feet in the area east of Bowman school (Barre quadrangle); south of this area it thins rather abruptly. The limestone is nowhere more than 15 feet thick and at most places it is much thinner. The tuff, on the other hand, varies to a remarkable degree, as shown in Figure 3. On the west side of the ridge between Roxbury and Bull Run, the breadth of outcrop of the tuff is approximately 1500 feet, and its true thickness is probably about 700 feet; it thins rapidly to the north and eventually disappears but it continues to the south as a unit 15 to 70 feet thick.

Distribution.—The Shaw Mountain formation is not exposed continuously, partly because it is comparatively thin, but nevertheless its horizon has been mapped from Bethel (Randolph quadrangle) northward to Hobart Mountain (Montpelier quadrangle), a distance of 42 miles along the strike, and traced in reconnaissance from Hobart Mountain to the Canadian boundary.

The trace of the Shaw Mountain formation can be followed south-southwestward along the higher parts of the ridge east of the North Branch of the Winooski River and through the western edge of the city of Montpelier. The formation crosses the Dog River 1.1 miles south-southeast of Montpelier Junction and extends southward along the west side of the river for a mile, then recrosses the river and appears on the east slope of the Dog River-Bull Run valley as far south as Cram Hill. South of this point it crosses a large, irregular highland area. Near Randolph it crosses the Third Branch of the White River. The trace of the formation coincides with Richardson's "great erosional unconformity," the boundary between argillaceous-arenaceous rocks on the west and the younger calcareous-argillaceous rocks on the east (Fig. 1).

Age.—Lenses of limestone were first discovered by Currier in 1936, at the lower Shaw Brook locality. He collected several specimens containing structures that strongly suggested crinoids but which were not certainly identifiable. Continued search by Jahns during mapping of the Barre quadrangle in 1938 resulted in the collection of fragments of undoubted crinoid stems. Since then additional collections were made at four localities, two of which are shown in Figure 4. According to Josiah Bridge (personal communication, 1938), the size and relative abundance of the columnals suggest that the beds cannot be older than

Middle Ordovician, and P. C. Raymond (personal communication, 1940) has concurred with this view. Edwin Kirk (personal communication, 1940) believes the crinoids may be even younger, although they are not generically or specifically determinable. Careful search in the field has thus far failed to reveal more than traces of any other fossil forms, but the positive evidence now at hand, though fragmentary and general, is of considerable value. On the basis of these crinoid stems, together with the presence of rocks of probable Middle Ordovician age stratigraphically below, the Shaw Mountain formation is tentatively referred to the Ordovician. Therefore, the so-called "Cambrian-Ordovician contact" of the earlier reports is at a horizon within the Ordovician system, and the base of the Ordovician lies somewhere to the west of the Shaw Mountain fossiliferous horizon.

The formation is underlain unconformably by the Cram Hill formation and overlain unconformably by the Northfield slate. No angular discordance is apparent along either unconformity.

NORTHFIELD SLATE

General statement.—The term Northfield slate was in common use by several of the earlier investigators of central Vermont geology. It never has been clearly defined stratigraphically, however, and additional confusion has arisen from use of the terms "Northfield conglomerate" and "Northfield phase of Irasburg conglomerate" (Richardson and Camp, 1919; Richardson, 1929a; 1929b; 1933) for the basal strata of the slate series. The Northfield slate, as here redefined, lies unconformably above the Shaw Mountain formation and conformably beneath the thick Waits River limestone of Richardson (1906). In general, it represents beds mapped in central Vermont by Richardson (1906) as "Memphremagog slates." However, the Memphremagog slates are not stratigraphically consistent as mapped from township to township and, moreover, they have been shown to be interbedded with limestone of the Waits River formation in some places, whereas in others they are clearly below the limestone. To eliminate some of the confusion that now exists as a result of past unorthodox terminology, the slates below the Waits River formation are here redescribed and redefined.

Lithology.—The Northfield slate is mainly light to dark gray. It is generally noncalcareous but contains scattered lenses and very thin beds of dense, bluish-gray limestone and dolomitic limestone. Like the Waits River limestone beds, the calcareous beds in the Northfield slate weather fairly deeply and develop a characteristic brown, spongy crust of leached material. The slates weather to a light gray, but some beds that contain

abundant ankerite porphyroblasts are covered with a light-brown stain.

Two prominent zones that contain many small siliceous nodules occur in the lower third of the section (Fig. 2). The lower of these zones is particularly well exposed in a large quarry 0.65 mile south-southeast of the Common in Northfield (altitude approximately 1140 feet). The nodules occur in close association with thin beds and lenses of sodarhyolite tuff and finely crystalline, blue-gray limestone. Tuffaceous beds are locally very prominent in the lower part of the slate east of Cram Hill.

A conglomerate that occurs at the base of the formation is significant. Thin and very poorly exposed, this conglomerate rests at several places upon the Cram Hill formation with what appears to be a slight erosional unconformity. The pebbles in the conglomerate have been drawn into thin elongate ellipsoids by tectonic action (Pl. 2, fig. 1), although many were doubtless somewhat tabular when deposited. They consist mainly of rock types found in the Cram Hill and Shaw Mountain formations—*i. e.*, gray to black quartzite, greenstone, sodarhyolite, and tuff. In addition, a pebble of blue crystalline limestone containing a crinoid columnal was found in the conglomerate on the saddle south of Ferry Hill (Randolph quadrangle). There can be little doubt that this pebble was derived from the crinoidal limestone horizon of the Shaw Mountain formation, a horizon that apparently is unique in central Vermont. This relation confirms the structural interpretation of the relative ages of the formations. The matrix of the conglomerate in most exposures is dark-gray slate, which is locally calcareous. A matrix of tan to light-gray siliceous limestone has been observed in two places.

This "Northfield conglomerate" or "Northfield phase of the Irasburg conglomerate" of Richardson has been recently found to represent a stratigraphic horizon distinctly different from that of the Irasburg conglomerate at its type locality (Richardson, 1906). For this reason the name Irasburg is untenable for the conglomerate at the base of the Northfield slate.

A group of thin, platy limestone beds lies directly above the conglomerate. These beds are lithologically similar to the limestone beds intercalated with the slates higher in the section, but the intercalated beds are not mappable and do not occur at consistent horizons. Variations in thickness and lithology of the conglomerate and limestone near the base of the section are shown in Figure 3. A locality list of the best exposures is as follows:

150 feet south-southwest of private crossing on Central Vermont Railroad, 1.2 miles south-southeast of Montpelier Junction (Conglomerate with abundant pebbles of volcanic rocks).

South bank of Dog River 0.15 mile downstream from railroad bridge, 2.5 miles southwest of Montpelier (conglomerate rich in quartzite and schist pebbles; *see* Figure 2 of Plate 2).

Lower part of hillside (altitude approximately 870 feet) 0.55 mile south-southeast of common in Northfield (limestone and conglomerate with limestone matrix).
Shaw Brook fossil locality (Fig. 4; type locality for "Northfield conglomerate"). The limestone crops out north of the brook 500 feet west of the conglomerate.
Cliff on northeast side of Cram Hill, 1.3 miles west of Bowman school (conglomerate).

Both conglomerate and limestone have been found as far north as Montpelier and as far south as Braintree Hill and they may extend even farther along the strike.

Thickness.—The methods that were used for determining the thickness of the Cram Hill formation were also employed in the study of the Northfield slate, although the lithologic homogeneity of the slate makes it difficult to trace folds and shear planes accurately on most exposures. A general factor of 0.5 to 0.55 was found to represent the relation between true thickness and breadth of outcrop and has been used for the determinations given in Figure 2.

The slate is thickest for about 10 miles along the strike between Northfield Falls and Belcher Hill (Barre quadrangle). It thins rather abruptly to the south, but the amount of thinning is somewhat obscured by shear zones and relatively large scale bedding-plane faults on the east side of its outcrop area. It is about 500 feet thick from Braintree Hill to a point 2 miles south of Randolph. South of this point it may pinch out altogether. The slate is also thinner north of Northfield Falls but does not seem to disappear entirely. For a distance of 12 miles in the Montpelier quadrangle it maintains a fairly consistent thickness of about 550 feet.

Distribution.—The Northfield slate is a generally resistant formation that has been traced in detail from Hobart Mountain (Montpelier quadrangle) to Braintree Hill (Randolph quadrangle), a distance of 32 miles. With few exceptions it forms prominent ridges wherever it is thick enough to have exercised appreciable control of the topography. The best example is the high, steep-sided ridge between Bull Run and Sunny Brook (Barre quadrangle), over most of which the slate is well exposed. The formation is named for the village of Northfield, near which the slate was quarried extensively many years ago. No continuous sections have been found, but the lithologic types are probably well represented along a westward-flowing brook on the south side of Turkey Hill (near Northfield) and along an eastward-flowing brook 1.7 miles northeast of Belcher Hill (Barre quadrangle).

Age.—Richardson originally correlated the Northfield slate with the slates at Montpelier and at Magog, Quebec, on the basis of lithologic similarities and later on the basis of paleontologic evidence (Richardson, 1902; 1906; 1916; 1919a; Richardson and Camp, 1919). He applied the

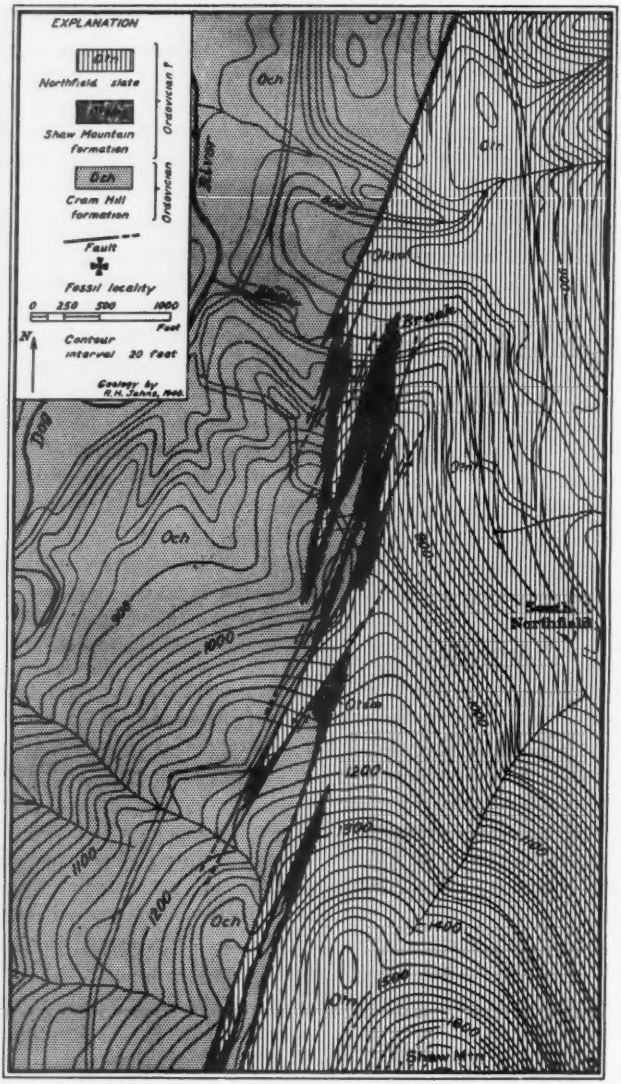

FIGURE 4.—*Detailed geologic map of a small area south of Northfield, Vermont*

All rocks dip 65° or more to the west; most of the dips are 80° or more. *See* Figure 1.

term "Memphremagog slate" to the slates at Northfield, Montpelier, and the localities to the north that were presumably the same formation as the slate at Magog. The paleontologic evidence upon which this correlation was based consisted of well-preserved graptolites found at Castle Brook, Magog, Quebec. These were identified as of Trenton age. Collections of "graptolites" were made later at Montpelier, Northfield, and other localities. These "graptolites" were obtained from thin-bedded limestones and calcareous phyllites of the Waits River formation as well as from the adjacent slate beds to the west. According to Richardson's reports, the specimens were identified by Rudolf Reudemann. The species reported ranged from lower Trenton to Beekmantown, and Richardson believed the slate beds to be younger than his Waits River limestone. This interpretation is conflicting, however, for the presence of supposed Beekmantown graptolites in the Northfield slate as reported later (Richardson, 1929b), p. 245) could hardly be reconciled with the lower Trenton age of some of the Waits River limestone beds, or of the Magog beds that were supposedly the equivalent of the Northfield-Montpelier slate horizons. The crinoidal limestone horizon described in the present report makes Richardson's interpretation of the age of the Northfield slate still more untenable.

Evidence that the Northfield slate is younger than the Shaw Mountain formation and older than the Waits River formation is compelling, for reasons set forth in this paper. It is therefore impossible to accept either the earlier interpretation of the relative ages of the formations or the verity of the "Cambrian-Ordovician boundary" at the base of the Northfield slates. Further, the Northfield slate and the fossiliferous Magog slates are now known to occupy different positions in the regional stratigraphic section.

This conflicting evidence, therefore, raised the question as to the validity of the "graptolites" reported from Richardson's Waits River limestone, Montpelier slate, and Northfield slates. The writers have made a careful examination of the slates at the reported fossil localities and of the many specimens in collections at Syracuse university. This investigation revealed no undoubted graptolites either in the exposures or collections. The rocks show, however, a profusion of markings that might easily be interpreted at first sight as organic forms. They are apparently similar in all respects to the markings seen in Richardson's collections and reportedly identified by Ruedemann. The markings appear to be caused by smears of fine micaceous minerals exposed as roughly parallel streaks upon cleavage planes. It is important to note that the cleavage planes cut the bedding planes at very sharp angles. With this fact established, the puzzling and suspicious phenomenon of general parallelism of the graptolites is explained. Foyles (1931) has

also noted the phenomenon. The writers have found no other markings that could be construed as graptolites, despite diligent search. The Beekmantown age of any of the formations above the Shaw Mountain formation is therefore discredited, particularly in view of the newly discovered crinoids and the present known structural relations. The Cram Hill formation as defined in this paper is held to be equivalent to part of the Magog slates, Trenton age, in Quebec, and a Lower Ordovician age for the Northfield slate is held to be untenable.

The Northfield slate is therefore tentatively correlated with the lower part of the Ordovician Tomifobia slates of Quebec (Clark, 1934). A further discussion of correlations with the formations in Quebec is given in a succeeding section.

It is pertinent to note at this point that several fossils other than graptolites have been reported from the Ordovician formations (chiefly limestone of the Waits River formation) of central and north-central Vermont as follows: Richardson (1902) cited a reference to an area of crinoidal limestone at Derby, Vermont, reported by C. H. Hitchcock. This area lies within the Waits River formation. Richardson also reported a crinoidal stem segment half an inch in diameter found in a block of Waits River marble, locality not stated but presumably in Orange County. Again (Richardson, 1929b, p. 245), he reported a possible crustacean, *Caryocaris*, from the Ordovician of Northfield; exact locality not stated. Finally (Richardson, 1939, p. 95-96), he reported another occurrence from Northfield (formation not stated) as follows:

> "Concerning the second sample Doctor Reudemann wrote, 'The coral-like body is certainly most suggestive. Corals are rare in these early Paleozoic rocks. It might more probably have been a cephalopod. It suggests in outline one of the many *Cyrtoceras* forms of the Trenton formation, but of course retains no septa and therefore is inconclusive.'"

These references concerning other fossils are cited to show that all available paleontologic evidence, other than graptolites, is consistent with the age interpretations set forth in this paper.

STRUCTURE

GENERAL STATEMENT

The formations described in this paper are included in a very thick series of Paleozoic beds that form the east limb of the Green Mountain anticline. The general trend of this structure is north-northeast. Steep westerly and vertical dips prevail so that the major structure is slightly overturned to the east. Steep easterly dips are to be found in local, narrow, discontinuous belts and are probably the result of drag folding and minor shearing on the limb of the regional anticline. There is no

significant repetition of the larger formational units across the general strike so that, so far as the Barre and adjacent quadrangles are concerned, successively younger formations are exposed from west to east.

The average dip decreases from west to east through the Barre quadrangle. In the western belt it is 80°-85°. In the eastern belt it is probably 55°-60°, though several dips as low as 40° have been noted. The general strike of the beds is N. 15°-20° E., and the local variations from this appear to be due mostly to drag folding.

That the formations are necessarily younger to the east and are on the east limb of an anticline that has a gentle northward plunge is demonstrable by the consistency of a great many observations made on drag folds, relation of cleavage to bedding, cross-bedding in some quartzite beds, and recognizable ripple marks in a few places. Many reconnaissance observations as far west as the axial ridge of the Green Mountains and more detailed studies along the valley of the Winooski and Lamoille rivers support the above interpretation of regional structure.

FOLDS AND FAULTS

Although the formations constitute an essentially unrepeated homoclinal series there are many local drag folds. These drag folds range from mappable size to minute crenulations, the larger ones being features of the more competent beds. Like the major regional structure to which they are genetically related, the drag folds for the most part plunge gently to the north. Other small folds related to horizontal regional shear of a later epoch of deformation plunge very steeply to the north.

The largest of the drag folds are found in the Shaw Mountain formation, and this localization is attributed to the relative competency of the formation as compared with adjacent formations.

No large thrust faults are known to exist in the mapped area, but a few minor thrusts were found. These indicate a slight and local overriding from the northwest.

The simplicity of the major regional structure, uncomplicated by large thrusts, has made it possible to trace the formations to the Canadian border with reasonable accuracy in reconnaissance fashion, on the basis of distinctive lithologic characteristics and sequences.

FOLIATION AND LINEATION

The strike of the regional foliation is essentially parallel to the strike of the bedding except on the local minor folds where the bedding departs from the regional trend. The dip of the foliation is, however, generally less steep than that of the bedding.

In the more homogeneous, less competent beds (slates and phyllites) the schistosity is so strongly developed that in many places it obscures the bedding. Even in the formations that contain interbedded quartzite layers the schistosity is generally as prominent as the bedding.

Two well-developed lineations are present: One of these pitches gently to the north and is a feature of the major regional structure; the other pitches very steeply to the north and reflects a later deformation. Lineations are chiefly recognizable by the alignment of porphyroblasts, elongation of pebbles in conglomerate beds, and the intersections of bedding planes with the foliation.

CORRELATIONS WITH FORMATIONS IN QUEBEC

As pointed out, Richardson's original correlation of the Northfield slate with the slates at Lake Memphremagog was based on lithologic features, and the interpretation of lower Trenton age was made later on the basis of the well-preserved graptolites found in the Magog slates at Castle Brook, Magog, Quebec. A conglomerate horizon at the base of the Northfield slates was correlated with a conglomerate found at Irasburg and named the Irasburg conglomerate. This bed was interpreted as a basal conglomerate of Beekmantown age, and strata below it were assigned to the Cambrian. Thus the beds that compose the Cram Hill formation of this paper were considered to be in the upper part of the "Missisquoi (Upper Cambrian) schists," and the crinoidal limestone at Shaw Brook (Shaw Mountain formation) was included in the Upper Cambrian strata, for Richardson (1919, p. 106) placed his Lower Ordovician (Beekmantown) boundary at the base of the conglomerate in the Northfield slate. This interpretation is now incompatible with the finding of crinoids in the limestone below the conglomerate horizon at this locality. The discovery by Jahns of a limestone pebble bearing a crinoid segment in an outcrop of the conglomerate at the base of the Northfield slate is highly significant.

The status of the Irasburg conglomerate is discussed in a succeeding section.

A study of Richardson's maps shows that his "Cambrian-Ordovician" boundary, or base of the slates that were traced southward to Northfield and beyond, connects with the eastern edge, or top of the Magog slates as mapped by Clark (1934). The lower Trenton fossil horizon of the Magog slates thus appears to be below the Northfield slate, a position that is compatible with the presence of crinoids in the Shaw Mountain formation at Northfield.

The lithologic sequence substantiates this interpretation. The Northfield slate is gray and siliceous; these are characteristic features that are

regional in their uniformity. The Magog slates are black, sooty, and much more carbonaceous. In the Cram Hill formation beds of carbonaceous slates are found, especially in the vicinity of Randolph and Hardwick, though these slates are thinner than the Magog beds. On the other hand, the lower part of the Tomifobia formation in Quebec (Clark, 1934) consists of beds that are lithologically like the characteristic Northfield slate and the associated highly individualistic limestone of the Waits River formation. The Magog slate, like the Cram Hill slates, is almost completely noncalcareous.

Clark (1934) interprets the contact between the Magog slates and the Tomifobia formation in Quebec to be a thrust fault in which the Magog slates have overridden the Tomifobia. He states that the general similarity of the graptolites in both formations may eventually lead to a correlation between them. However, the presence or absence of this fault and the possible age equivalency do not affect the general interpretation of this paper, for the important correlation is concerned chiefly with the Magog slates. The data in the Barre quadrangle, however, would suggest that the Tomifobia slates and limestones are generally younger than the Magog slates, both probably being of Ordovician age.

The suggested correlation is therefore as follows:

	Central Vermont	Southern Quebec (Clark)
Ordovician (?)	Northfield slate	Basal parts of Tomifobia formation
	Shaw Mountain formation (crinoidal)	(Horizon of "Bunker Thrust")
Ordovician (Middle)	Cram Hill formation	Upper parts of Magog slates

STATUS OF THE IRASBURG CONGLOMERATE

The writers have visited the type locality of the Irasburg conglomerate and consider it to be an intraformational bed within the Waits River formation and therefore not to be correlated with the base of the Ordovician system nor with the Magog conglomerate of Dresser (1925). No conglomerate analogous to the Magog conglomerate has yet been found in the course of detailed mapping in the Barre and adjacent quadrangles, although it may ultimately be found stratigraphically below the Cram Hill formation.

The Irasburg conglomerate therefore appears to be without important regional significance as a stratigraphic unit.

CONCLUSIONS

The discovery of crinoidal limestone in the beds below the so-called "Cambrian-Ordovician contact" of central Vermont necessitates a revision in the correlation of the beds above and below this boundary. Not only do crinoids appear to be unknown anywhere in rocks of Cambrian age, but the size and degree of development indicated by the fossils found in the Shaw Mountain formation strongly suggest a Middle Ordovician or later age for this formation, according to the opinions of Bridge, Kirk, Raymond, and Ulrich. Accordingly the Northfield slate and succeeding beds must necessarily be interpreted as of Middle Ordovician or later age, and Richardson's "Cambrian-Ordovician contact" becomes a horizon within the Ordovician system. This correlation is apparently in harmony with the sequence, lithology, and fossils of the Magog slate and Tomifobia slates and limestones of Quebec. The earlier confusion arising from the supposed graptolites of the Northfield slate and Waits River formation as well as the conflicting interpretations of the structural relations and relative ages of these beds thus seem to be clarified. All available data thereby become mutually supporting as consistent evidence.

The position of the actual Cambrian-Ordovician boundary is not known. It apparently lies somewhere to the west of the Cram Hill formation—that is, the upper part of the "Missisquoi group" of Richardson, which he believed to be Upper Cambrian. It may lie west of this entire group, but, so far, detailed mapping has not revealed an erosional or lithologic break that can be satisfactorily interpreted as representing the boundary. There may be no sharp break between Cambrian and Ordovician strata, for it is well known that in fossiliferous areas in the northeastern United States a strong unconformity at this horizon is rarely encountered.

It is suggested that the unconformity at the base of the Shaw Mountain formation may well be analogous to that at the base of Kay's Shoreham member of the Sherman Fall formation (Kay, 1937, p. 264-271) on the east side of the Adirondack arch in New York. According to W. M. Cady (personal communication, 1941), this unconformity, which occurs in the lower part of the middle Trenton, is a widespread feature; it extends across the Champlain Valley and as far eastward as the western flank of the Green Mountains in the latitude of Rutland, Vermont. Such assignment of the Cram Hill-Shaw Mountain unconformity to the early middle Trenton is in full accordance with all available data, and in particular is compatible with the reported Trenton graptolites in the lower Tomifobia slates of Quebec (Clark, 1934), with which the Northfield slate is correlated. Further work needs to be done on both the

Waits River and Tomifobia formations, but the succession of lower beds and the interpretation of their age relationship as outlined in this paper seem well substantiated by the data now at hand.

It is not surprising that the interpretation of Beekmantown graptolites in the Northfield slate and Waits River limestone beds should have led to the earlier view, for in Vermont there was no other paleontologic evidence to conflict with it. However, the discovery of a crinoidal limestone bed stratigraphically below the Northfield slate and Waits River formation makes necessary a revision of geologic chronology in central Vermont. The first step in the revision has been presented in this paper. It should be stated that the extensive reconnaissance work of Richardson has not only been of great value in tracing the broader lithologic facies of the lower Paleozoic strata throughout the central part of the State, but has provided a helpful basis on which to build the current work. Such pioneer work in a region long avoided by others is to be commended for the general information and stimulation for further and more detailed work that it provides.

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